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Abstract

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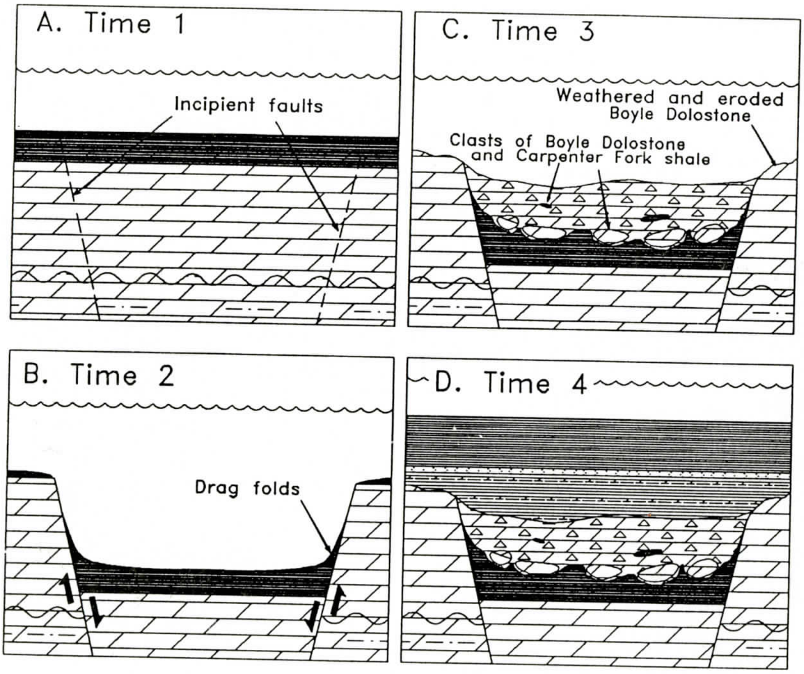

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DOCUMENTATION AND EVALUATION OF RADIOCARBON DATES FROM THE "CAPE FEAR COQUINA" (LATE PLEISTOCENE) OF SNOWS CUT, NEW HANOVER COUNTY, NORTH CAROLINA

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ABSTRACT

Conventional radiocarbon analysis of three shallow marine shell samples, each of a different taxonomic species, suggest that the Cape Fear coquina, which is now positioned at or slightly above modern mean sea level, was deposited around 24 ka to 32 ka B.P. The hypothesis that these apparent radiocarbon ages are the result of contamination of much older, circa 124 ka B. P., material by younger material was found to be unsupported. The only identified contaminant are post depositional calcite cements which in the worst possible case could not have exceed 0.25% of the volume of the samples. These same calcite cements were also analyzed by conventional radiocarbon methods and were found to present an apparent mean age of cement formation of 26 ka B. P. The oldest possible apparent radiocarbon age for the depositional event of the Cape Fear coquina is calculated to be 30 ka B. P. while the mean age of the cementation event is 18 ka B. P.

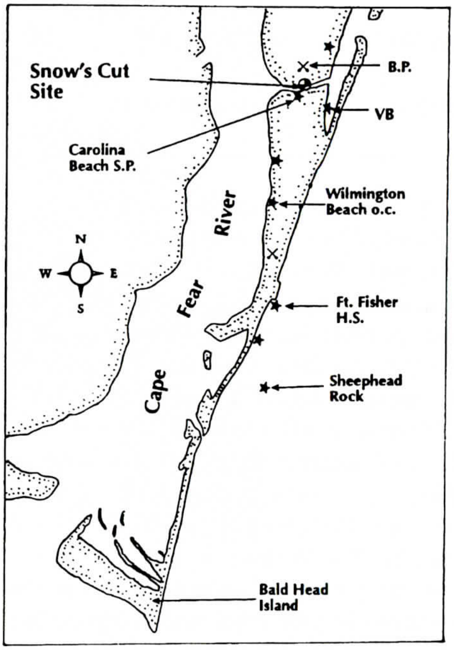

Figure 1. Location of Snow's Cut Site and other exposures of the "Cape Fear coquina". "B.P." denotes the Burnett Pit; "VB" is a vibracore location; crossed shovels denote marl or burrow pit utilizing the coquina; and stars denote other significant exposures of the coquina examined in this study.

INTRODUCTION

Four radiocarbon dates (Table 1), three from marine mollusks shells and one from post depositional calcite cements were obtained from a Late Pleistocene shallow marine to intertidal deposit of the "Cape Fear coquina" found at Snows Cut (34^{o} 3' 14" N, 77^{o} 54' 23" W) 20 km south of Wilmington, North Carolina (Figure 1). The apparent radiocarbon ages of the mollusk samples exhibit a range from 24 ka to 32 ka. These apparent radiocarbon ages present a problem in that they are inconsistent with other estimates of the age of the "Cape Fear coquina". Prosser (1993) applied the uranium series method to shells of *Mercenaria* sp. collected at Snows Cut and obtained ages ranging from 53 ka to 70 ka (Table 2). Wehmiller and others (1988), applying amino acid racemization, found a mean D/L Leu value of 0.52 from two specimens of *Mercenaria* sp. which were collected at Snows Cut. Such a value corresponds to amino zone IIId of Wehmiller and

Table 1. Radiocarbon dates from the "Cape Fear coquina".

Lab Number	Conventional Age	Material dated	Location
Beta-37432	24,390±420 B.P.	60 *Nassarius obsoleta* shells.	Snows Cut
Beta-37433	32,150±550 B.P.	325 *Donax variabilis* valves.	Snows Cut
Beta-37434	26,380±500 B.P.	750 'dog tooth' calcite spar crystals filling *Busycon* sp. and *Mercenaria* sp. shells	Snows Cut
Beta-38472	29,000±700 B.P.	*Crassostrea virginica* cluster	Snows Cut
Beta-27840	31,670±515 B.P.	*Crassostrea virginica* (William_Cleary, personal communication)	Carolina Beach NC
Beta-37431	34,320±960 B.P.	940 'dog tooth' calcite spar crystals filling *Busycon* sp. shells	Ft. Fisher Historic Site
unknown	39,040±1645/-2070 B.P.	*Mercenaria* sp. shell (Victor Zullo personal communication).	Burnett Pit

others (1988) which apparently has an age in excess of 220 ka. A more recent work by Wehmiller and others (1995) notes the additional amino acid analysis of 16 specimens of *Mercenaria* sp. from Snows Cut. The A/I values of these "cluster into two apparent aminozones with mean values of approximately 0.46+/-0.025 (n=4) and 0.34+/-0.025 (n=12)" (Wehmiller and others 1995, p.331). The lower ratio corresponds to isotope stage 5, the higher ratio to stages 7 to 9 (Wehmiller and others, 1995, figure 2). The Wehmiller and others (1995) paper unfortunately does not present any documentation on these shell samples from Snows Cut, however it does report documentation on approximately 200 shells from the region. Of particular interest is the color of the *Mercenaria* sp. shells; their Table 22 notes a number of specimens with a purple coloration. The A/I ratio reported there for the shells with the purple coloration ranges from 0.0 (modern) to a maximum of 0.24. These purple colored shells therefore all appear to be younger than isotope stage 5. The Snows Cut site also contains *Mercenaria* sp. with purple coloration. This suggests a maximum age for the Snows Cut depositional event as being less than the age of isotope stage 5. Sample JW93-09-06 of Wehmiller and others (1995), which had the purple coloration, was also radiocarbon dated (Beta 62757 @29,900+/-290; AA-11807 @ 43,100+/-1200). This also suggest a less than isotope stage 5 age for the samples though the difference between the two age determinations for sample JW93-09-06 may indicate sample contamination problems. A single *Mercenaria* sp. from Stetson Pit, Dare County, North Carolina was reported by (Wehmiller and others 1988) to have a D/L Leu value of 0.33. A coral from the same site yielded a uranium series age of 75 ka (Szabo, 1985) thus suggesting the Snows Cut material to have a minimum age in the neighborhood of 75 ka. The oxygen isotope based sea level curve of Chappel and Shackleton (1986), when applied strictly without allowances for crustal adjustments to post glacial conditions, indicates that the Snows Cut material could have been deposited no later than early stage 5, ~120-130 ka B.P.

Table 2. Uranium series apparent ages of *Mercenaria mercenaria* shells collected at Snows Cut section (from Prosser, Table 2, 1993).

Sample Number	Apparent Age
S.C. 2	18±0.5 ka
S.C. 21	57±3 ka
S.C. 22	65±3 ka
S.C. 23	57±4 ka
S.C. 24	68±5 ka
S.C. 25	62±4 ka
S.C. 26	65±4 ka
S.C. 27	73±7 ka

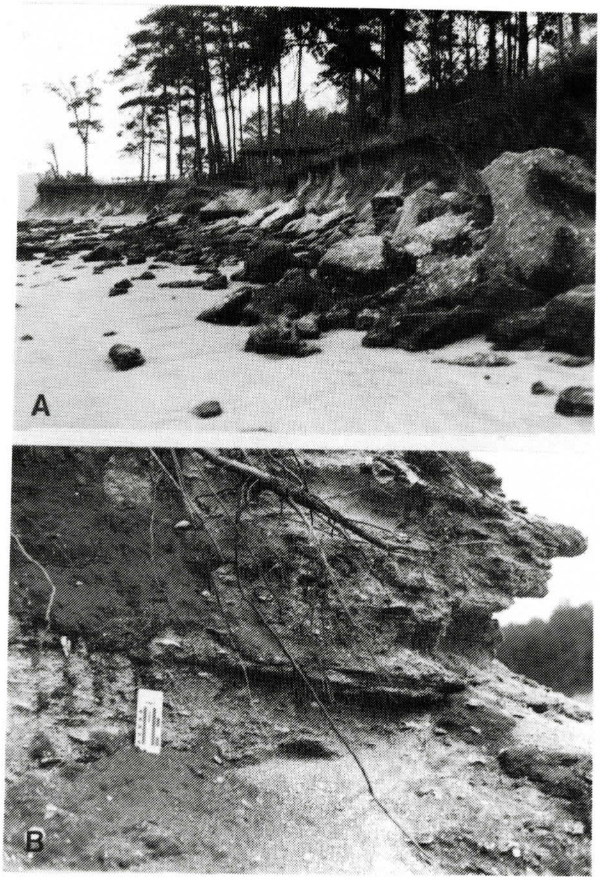

Figure 2. Outcrop photographs of the Snows Cut Site. (A) is a view looking west along the north wall of Snows Cut. Intracoastal Waterway is to the left. Bank below fence at left center of the photograph is the location of the shell hash from which the radiocarbon samples were collected. (B) is the actual portion of the bank from which the samples were obtained. View is looking eastward. Photograph was taken three years after the samples were collected.

Samples which give radiocarbon ages in excess of 20 ka and which are inconsistent with other age determinations are almost always written off without elaboration or qualification as having been contaminated by young carbon. This is unfortunate in that such samples may actually be uncontaminated and thus capable of providing clues about the reliability of the other methods of dating. Even if they really are contaminated these samples may provide clues about the contamination process. The purpose of this paper is to provide a comprehensive documentation of the radiocarbon dated material from Snows Cut and to evaluate the possibility that these samples are in some way contaminated.

The significance of this study is that the Snows Cut site represents one of approximately two dozen sites worldwide where the radiocarbon method and at least one other dating method have been applied to the same or spatially related material which appears to have an absolute age greater than 25 ka (see Dockal 1995). Of those two dozen or so sites only 10, including the Snows Cut site, have both methods of dating applied to either the same sample or to two different samples from the same sediment layer. It is well established that reported

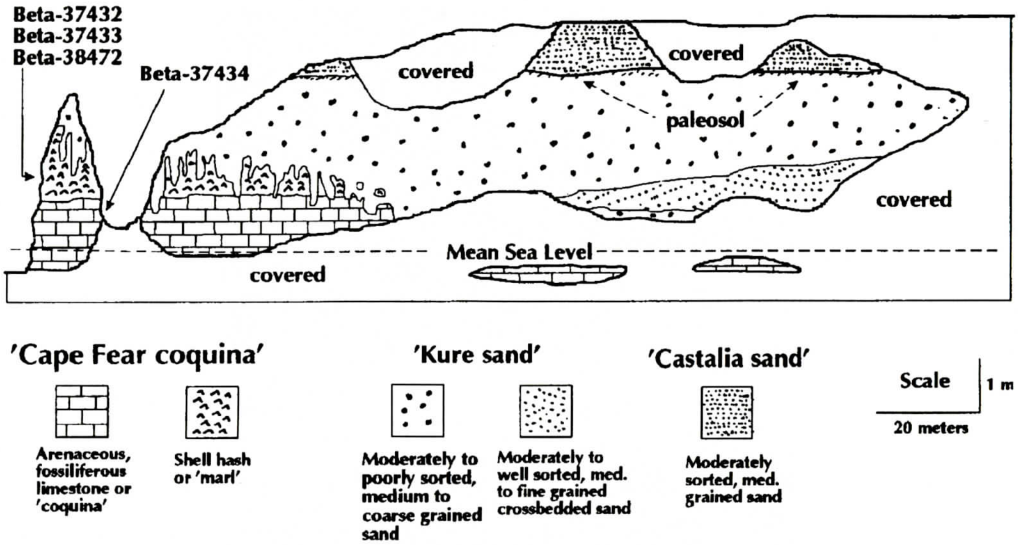

Figure 3. Diagram of the north bank of Snows Cut illustrating the relationship of the various lithologic units. Vertical exaggerations 20X.

radiocarbon dates are *apparent* dates that need to be calibrated to bring them in line with absolute time. This is primarily due to the fact that the rate of radiocarbon input into the atmosphere, hydrosphere, and biosphere is not constant. For dates less than approximately 10 ka B.P. dendochronology is applied to the calibration of the radiocarbon time scale. For dates up to 25 to 30 ka B.P. calibration has been achieved using U-Th ages of corals (Bard *et al.* 1990). Beyond that range there are as yet too few published and documented dates to attempt a sound calibration.

The reader is cautioned to keep in mind that the terms "absolute age" and "apparent radiocarbon age" are not the same though both are expressed as years before present. The apparent radiocarbon age of a sample expresses the *assay amount* of radiocarbon found within the sample. If and only if the sample is free of contamination from older or younger carbon, the cosmic ray flux at the time of sample formation was equal to the modern flux, and a starting point of the year 1952, is taken can one directly relate "apparent radiocarbon age" to "absolute age" (Libby, 1955). In the text that follows always assume unless otherwise noted that any expressed ages are "apparent radiocarbon ages" and that the concept they reflect is that of a radiocarbon assay concentration. Stating that a sample has an age of 27 ka B.P. therefore does not mean that the sample is 27,000 absolute years old; it does however mean that the sample has the same concentration of radiocarbon within it that a similar sample would have had if it formed 27,000 absolute years before 1952, formed under a cosmic ray flux equal to that of the modern flux, and is absolutely free of older or younger carbon.

DOCUMENTATION

The samples collected for radiometric dating were taken from the north side of Snows Cut, an artificial entrenchment made about 1930 to connect the Intracoastal Waterway to the Cape Fear River. The sample site is located 600 meters west of the US 421 highway bridge over the waterway and approximately 2.2 kilometers west of the modern Atlantic beach (Figure 2). This site is the same as "location 24" of Carter and others (1988). Material was collected within weeks of a major winter storm

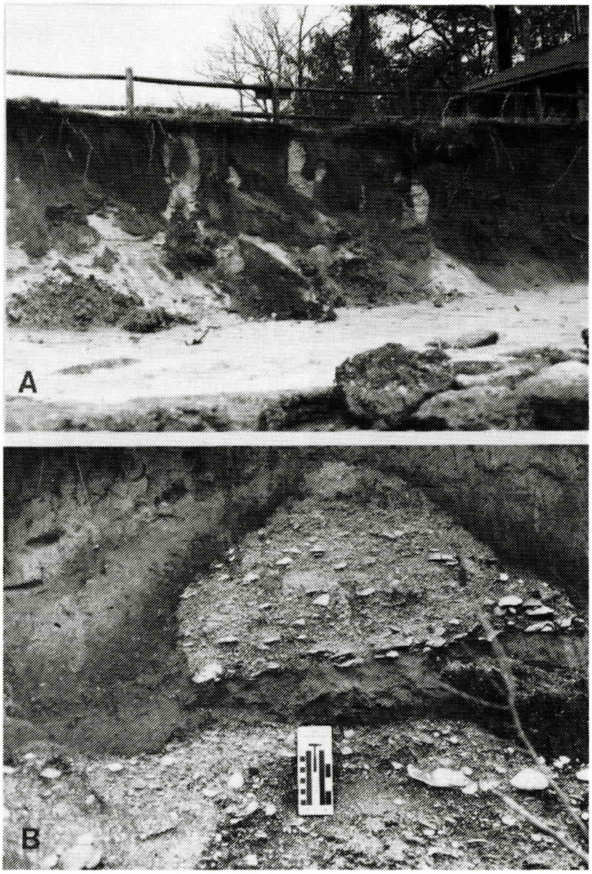

Figure 4. Photographs illustrating the boundary relationship between the Cape Fear coquina and the Kure sand at Snows Cut. (A) Fingers of light colored shell hash within the darker poorly sorted sands of the Kure sand. (B) A pseudolithoclast of shell of unlithified shell hash surrounded by typical Kure sand.

which caused the partial collapse of the north bank of Snows Cut and exposure of fresh unweathered material.

Host Lithology

All the samples were collected from an informal stratigraphic unit known as the Cape Fear coquina. The term "Cape Fear coquina" was first applied by Wells (1944) to Pleistocene coquina occurrences in then neighborhood of Snows Cut and Fort Fisher (Figure 1). The term is now applied to any non-cemented to poorly cemented fossiliferous Pleistocene bed in New Hanover County, North Carolina and in the adjacent offshore waters. On the north side of Snows Cut the Cape Fear coquina consists of two lithofacies; an unlithified, sandy, shell hash lithofacies which overlies a poorly cemented, friable, arenaceous limestone lithofacies (Figure 3).

The shell hash lithofacies is characterized by weak grain size sorting, weakly marked trough and tabular planar cross stratification, and a strongly developed sense of shell imbrication with the shells assuming a concave down orientation. Cross stratification and shell imbrication indicate a southerly sediment transport direction. The sediment consists of 60% shell debris, 39% quartz or other siliciclastic grains, and one to two percent lithoclasts (Table 3). The majority of the shells are fragmented, well rounded, and generally unidentifiable. In thin section these appear typically micro bored with

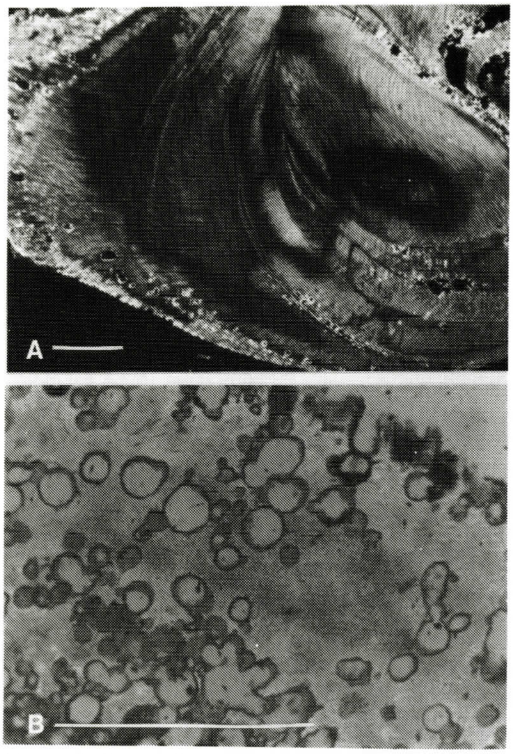

Figure 5. Photomicrographs of microborings in bioclasts taken from the unlithified shell hash at Snows Cut. (A) *Nassarius obsoleta* and (B) unidentified rounded shell fragment. Bars = 0.5 mm.

the borings remaining completely open and not occluded by cements or mud filling (Figure 5A and 5B).Some of the larger shells and shell fragments, which have concave down surfaces within the strata, have minor amounts of pendant or stalactitic calcite cement developed within the shelter porosity (Figure 6A). A few of the shell fragments appear slightly etched. Permineralization of echinoderm fragments is completely absent. The lower portion of the shell hash lithofacies has weak streaks or zones of cementation which follow bedding and zones of finer size sorting. These first appear at about the midpoint of the outcrop and increase in frequency downward. This cementation is due to a meniscus cement made up of small, equant calcite crystals (Figure 6B). Within these same zones the etching of bioclasts is more prevalent and the development of moldic porosity after bioclasts is evident. Permineralization of echinoderm fragments is evident while the infilling of the micorborings is absent, even those immediately adjacent to the meniscus cement. The volume of calcite cement is slightly greater than the volume of dissolved carbonate (moldic porosity) (Table 3). In weathered outcrop the boundary between the overlying unlithified shell hash and the underlying limestone appears abrupt and undulate with nearly two meters of relief. Owens (1989) interpreted this abrupt change to be an unconformity separating his overlying Wando Formation from the underlying Waccamaw Formation. However, this change in lithology is not depositional but instead it is a change in the diagenetic character of the strata. Separating the two lithofacies is a transition zone where the unlithified shell hash becomes

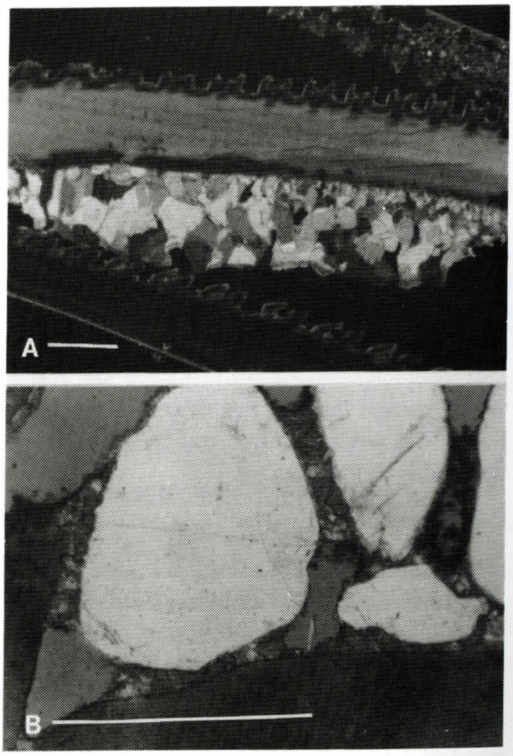

Figure 6. Cements and diagenetic features from the weekly lithified portion of the shell hash (A) Pendent cement within the shelter porosity developed below a *Donax variabilis* shell. (B) Meniscus cement. Bars = 0.5 mm.

Table 3. Modal analysis of typical samples of the "Cape Fear coquina".

	SCN22	SCN24	SCN20	SCN21	SCN25	SCN26	SCN27	FF53	FF56	WB10
	Snows Cut, Intercoastal Waterway							Ft. Fisher HS		Wil-mington Beach
	Non Cemented Layer		Weakly Cemented Layer		Moderately Cemented Layer			Mod. Cmt. Layer		Mod. Cmt. L.
A. Aragonitic Bioclasts	58±8	57±8	24±6	25±5	5±3	8±6	12±5	8±4	9±4	8±4
B. Calcitic Bioclasts			<1	1	<1		9±4	<1	1	
C. Calcitized Aragonitic Bioclasts	n.p.	n.p.	n.p.	n.p.	9±4	6±4	6±3	2±2	5±3	15±5
D. Detrital Quartz & Feldspar	39±9	37±8	42±6	42±6	26±5	34±9	40±7	34±7	32±6	16±5
E. Rock Fragments	1	2	<1	1						
F. Interpartical Porosity	n.a.	n.a.	16±4	22±5	10±4	7±5	2	13±4	8±4	1
G. Intrapartical Porosity	2±1	4±3	1±1	2±1						
H. Moldic Porosity	n.p.	n.p.	6±3	2±1	25±5	25±9	16±5	25±6	26±5	30±6
I. Calcite Cement	n.p.	n.p.	10±4	6±3	24±5	20±8	16±5	18±5	19±5	30±6
TOTAL	100	100	99	101	99	100	100	100	101	100
(1) Present Total Porosity	n.a.	n.p.	23±5	26±5	35±6	32±10	18±5	38±7	34±6	31±6
(2) Intial InterparticalPorosity	27±2 (8)	27±2 (8)	27±6	30±6	34±6	27±9	18±5	31±6	27±6	31±6
(3) Intial Bioclasts	44 (9)	45 (9)	31±7	34±5	39±7	39±12	43±9	35±7	41±7	53±9
(4) RatioMoldic porosity to Cement	n.a.	n.a.	0.6	0.3	1.0	1.3	1.0	1.4	1.4	1.0
(5)Ratio: Bioclasts to Quartz & Feldspar	1.5	1.6	0.7	0.8	1.5	1.1	1.1	1.0	1.3	3.3
(6) Total CaCO3	n.a.	n.a.	34±7	32±6	38±7	34±11	43±9	28±7	34±7	53±9
(7) Ratio:Present to Intial Porosity	n.a.	n.a.	0.9	0.9	1.0	1.2	1.0	1.2	1.3	1.0

(1) F+G+H
(2) F+I
(3) A+B+C+G+H
(4) H/I
(5) (A+B+C+G+H)/D
(6) A+B+C+I
(7) (F+G+H)/(F+I)
(8) Weighted mean from other thinsections
(9) Calculated assuming intial interpartical porosity

increasingly downward more lithified by meniscus calcite cement

The meniscus cement then becomes progressively obscured by bladed calcite cement which in turn becomes blocky calcite cement with a poikiotopic texture. This later cementation character is typical of the limestone lithofacies. With this downward change in cement character there is a parallel increase in the etching of the bioclasts and frequency and quantity of moldic pores after aragonite bioclasts (Table 3). Calcite bioclasts exhibit no dissolution features. This transition zone is not flat lying but instead has a recognizable slope to it and cuts across traceable bedding features. A horizon of larger shells and shell fragments, about 10 centimeters thick, was noted near the west end of the exposure site in 1990. This shell zone could be traced from a point where it was entirely within the limestone lithofacies, to where it passes through the cementation transitional zone, and on into the shell hash lithofacies. This occurred over a lateral distance of 10 meters. The cementation transition zone clearly cuts across bedding features thus indicating it is a post-depositional diagenetic feature and not a depositional phenomena.

Within the limestone lithofacies permineralization of echinoderm fragments is ubiquitous. Moldic pores generally lack later calcite cementation. Microborings are either still open or completely filled with calcite spar. Modal analysis of typical samples indicates the volume of moldic porosity is equal to or slightly greater than the volume of calcite cement present. Aragonitic bioclasts which have not been lost due to dissolution are partially or completely replaced by calcite (calcitization of aragonite) (Figure 7A). In all other characteristics the two lithofacies are alike. Both are here interpreted to be of the same original depositional lithology and to have been deposited at the same time during a single depositional event.

Rock fragments represent up to one to two percent of the clasts of the Cape Fear coquina (Table 3). They are chiefly a calcite cemented, fine grained, subangular, well sorted, quartz

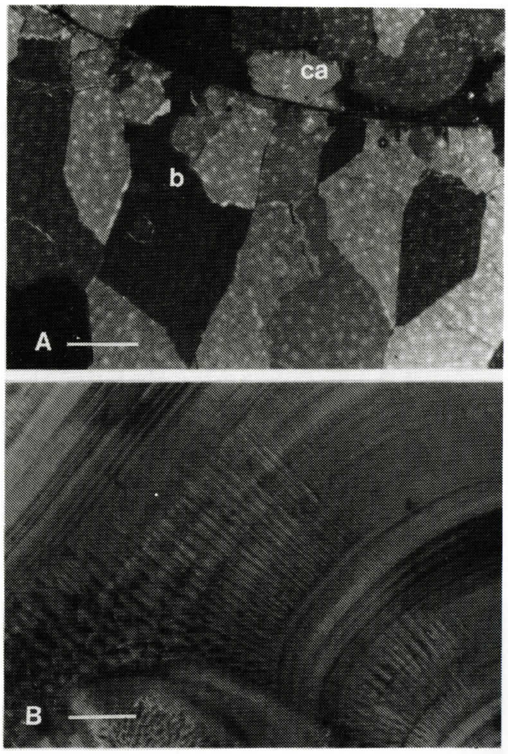

Figure 7. (A) Bladed porosity filling calcite cement "b" and calcitized aragonite "ca" of a *Busycon* sp. shell from the lithified portion of the Cape Fear coquina at Snows Cut. Spotted texture is an artifact of staining with alizarin red-s. (B) *Crassostea virginica* (Beta-38472). All with cross polarizers. Bars = 0.5 mm.

arenite which is lithologically similar to the Pee Dee Formation (Cretaceous) as exposed in the Castle Hayne quarries 35 kilometers to the north. A single abraded and slightly rounded rock fragment containing the oyster *Conradstrea lawrencei* was found in the shell hash lithofacies. This oyster is characteristic of the Waccamaw Fm. (Pliocene-Pleistocene) and its occurrence here within a lithoclasts indicates that the Cape Fear coquina postdates the Waccamaw. Several rock fragments ranging up to 15 centimeters in diameter of a coquina like lithology were also found within both lithofacies of the Cape Fear. Together with the Waccamaw fragments these rock fragments indicate the presence of reworked older material within the Cape Fear. Discoidal quartz pebbles one to three centimeters in diameter and prolate to spherical, rounded to subangular, granules and pebbles of several lithologies attributable to the Piedmont Province can also be found within the Cape Fear coquina and adjacent Kure sand.

The coquina at Snow's Cut contains *Anadara brasiliana* (Lamarck), a mollusk which is characteristic of Blackwelder's (1981) upper Pleistocene and Holocene Yongesian Substage of the Longian Molluscan Stage. Other identified mollusks collected in place from Snows Cut and Ft. Fisher are listed in Table 4. Two corals can be found in the coquina, *Septastraea crassa* and *Siderastrea radians.* The former, which is common in the Waccamaw Fm., occurs here only as abraded fragments while the latter was found encrusting specimens of *Mercenaria* sp. and without signs of abrasion. The barnacle *Balanus improvisus* (see Zullo and Miller 1986) can be found attached to *Crassostrea virginica* or as individual plates. Fragments of sand dollars, probably *Millita* sp. plus forams, crab claws, fish teeth, and an occasional well rounded bone fragments are also found within the coquina.

The majority of the mollusk shells are fragmented and abraded to varying degrees. However, specimens of *Nassarius trivillatus*, *Tagelus plebius*, *Polinices immaculatus*, *Sinum perspectivum*, as well as *Balanus* sp. exhibit no abrasion and specimens of *Donax variabilis*, *Anomia simplex*, and *Brachidontes recurvus* are only occasionally abraded. All specimens of *Aequipectin gibbus*, *Anardra ovalis*, *Glycerimus pectinata*, *Noetia ponderosa*, *Mercenaria mercenaria,* and *Busycon contrarium* exhibited extensive abrasion. The abrasion of all other molluscan species ranged from none to being well rounded. Approximately 5% of the specimens of *Mercenaria mercenaria* retained the purple coloration in the area of the ventral margin and pallial sinus. Most of the specimens of *Aequipectin gibbus* had some reddish, purple and yellow coloration. *Nassarius obsoletus* also had a brown coloration. Articulation of the bivalves was completely absent. Very few mollusk shells were drilled by predators, however over half of the rounded shell fragments observed in thin section have been extensively microbored.

The molluscan fauna of the Cape Fear coquina suggest deposition under climatic conditions similar to that found today along the southeastern Atlantic seaboard between 34 and 36 degrees north latitude (Figure 8). The fauna

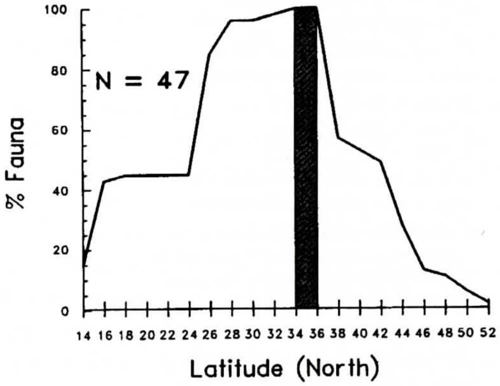

Figure 8. Percent of the molluscan fauna of the Cape Fear coquina at Snows Cut that are found alive today at various latitudes along the Western Atlantic coast. Shaded area represent the best match; 34^o to 36^o degrees North. Forty seven (N=47) taxa were evaluated.

represent a mixture of taxa from several environments; predominately salt marsh, beach surf zone or intertidal, and shallow shelf (water depth less that 10 meters). The presence of the clam *Rangia cuneata* indicates a close proximity to a fluvial system. The lithoclasts with a Piedmont provenance further suggests a close proximity to fluvial processes. However, some of the lithoclasts had been within the surf environment long enough to have developed a well rounded and discoidal shape. The size of the bioclasts and lithoclasts of the coquina together with the cross-bedding and imbrication of the shells are interpreted to indicate deposition within a high energy environment. Deposition may have been the result of a powerful storm. Similar but smaller shell debris deposits were noted to have formed on the area's beaches after Hurricanes Gloria and Diana. However,

Table 4. Mollusca from the Cape Fear Coquina New Hanover Co., North Carolina

	Snows Cut non-cemented portion *	Ft. Fisher Historic Site *
Mollusca: Pelecypoda		
Aequipectin gibbus (Linnaeus)	common	rare
Aequipectin irradians (Lamarck)	rare	common
Anadara brasiliana (Lamarck)	common	
Anadara (Lunarca) ovalis (Bruguiere)	abundant	rare
Anadara transversa (Say)	common	
Anomia simplex (Orbigny)	common	
Brachidontes recurvus (Rafinesque)	rare	
Chione cancellata (Linnaeus)	very rare	
Crassoatrea virginica (Gmelin)	abundant	abundant
Cyrtopleura cosata (Linnaeus)	rare	
Dionocardium robustum (Lightfoot)	common	common
Donax variabilis (Say)	abundant	abundant
Dosinia discus (Reeve)	rare	
Ensis directus (Conrad)	rare	rare
Glycerimus americana (Defrance)		rare
Glycerimus pectinata (Gmelin)	rare	
Labiosa plicatella (Lamarck)	rare	
Mercenaria campechiensis (Gmelin)	abundant	
Mercenaria mercenaria (Linnaeus)	abundant	abundant
Mulina lateralis (Say)	common	
Noetia ponderosa (Say)	abundant	common
Petricola pholadiformis (Lamarck)	common	
Rangia cuneata (Gray)	abundant	rare
Spisula solidissma (Dillwyn)	rare	common
Tagelus plebius (Lightfoot)	common	rare
Trachycardium muricatum (Linnaeus)	rare	
Venericardia tridentata (Say)	common	
Mollusca: Gastropoda		
Busycon contrarium (Conrad)	common	common
Busycon canaliculatum (Linnaeus)	rare	
Crepidula convexa (Say)	common	
Crepidula fornicata (Say)	rare	
Crepidula plana (Say)	rare	
Epitonium novangliae (Couthany)	very rare	
Eupleura caudata (Say)	rare	
Nassarius albus (Say)	very rare	
Nassarius obsoletus (Say)	abundant	
Nassarius trivattatus (Say)	common	
Oliva sayana (Ravenel)	very rare	
Olivella mutica (Say)	rare	
Polinices duplicatus (Say)	common	rare
Polinices immaculatus (Totten)	rare	
Prunum apicinum (Menke)	common	
Retusa candei (d'Orbigny).	rare	
Serpulorbis decussata (Gmelin)	very rare	
Sinum perspectivum (Say)	very rare	
Terebra dislocata (Say)	rare	
Urosalpinx cinerea (Say)	rare	

*After several hours of working the outcrop one would have found: *'very rare'* one specimen; *'rare'* two to ten specimens; *'common'* more than 10 but less than 100 specimens ; or *'abundant'* more than 100 specimens.

these deposits disappeared within a few weeks due to reworking during high tide. A very large storm could have resulted in a shell debris deposit above mean high tide and thus unlikely to have been later eroded or reworked by later wave action. This would be especially true if deposition occurred during a period of declining global sea levels. The lack of articulation of the bivalves and the lack of borings by marine predators further points to subareal post deposition desiccation of storm accumulated mollusks

The Cape Fear coquina is overlain at most sites by a medium to coarse grained, moderately to poorly sorted, near symmetrical to coarse skewed, platykurtic to leptokurtic, weakly indurated, ferruginous sublitharenite (*in sensu* Folk, 1980). Sedimentary structures are absent in this sand in the area from which the radiocarbon samples were collected. Elsewhere this sand does exhibit some cross bedding which is limited to lensoidal pockets. Wells (1944) referred to this as the "Kure sandstone" while Moorefield (1978) referred to it as the "Unit I Bluff Sands".

At Snows Cut the Kure sand can be seen forming fingers that extend downward into the shell hash lithofacies and occasionally into the limestone lithofacies (Figure 4A). These are as much as 1.5 meters in length and 10 centimeters in diameter. The central core of each finger is free of carbonate grains and cement. Grain size distributions of the material within the fingers and that of the immediately adjacent insoluble fraction of the coquina are identical in mean, mode, standard deviation, skewness, and kurtosis. Within the Kure sand at the east end of the Snows Cut site are found what would appear to be lithoclasts of the unlithified shell hash lithofacies (Figure 4B). There is no conceivable way that these could have been transported and thus probably represent residual pockets of carbonate material surrounded by insoluble material. Similar lithoclasts or actually pseudolithoclasts of lithified coquina up to one meter in diameter were also found within the Kure sand near Wilmington Beach (Figure 1). Others are commercially gathered from throughout the area for use as landscaping stones. Also at the east end of the coquina exposure at Snows Cut a layer of larger shells and shell fragments, coarse sand, and pebble lithoclasts was observed about half way up the non-cemented portion of the coquina. This, when traced laterally eastward, cuts across the boundary between the shell hash and the Kure sand where it continues on as a zone of coarser sand and pebbles.

The character of the fingers of Kure sand cutting into the coquina, the existence of pseudolithoclasts of coquina within the Kure sand, and the traceability of pebble marker horizons across the boundary between the coquina and Kure sand all indicate that the Kure sand was originally Cape Fear coquina. The Kure sand is simply the insoluble residue that remained after the carbonate fraction of the Cape Fear was removed by dissolution. The cross-cutting relationship of the fingers of Kure sand and the boundary between the two lithofacies of the Cape Fear coquina indicate that cementation of the coquina preceded the dissolution development of the Kure sand.

At the top of the Kure sand is a well defined paleosol which forms a smooth, flat-lying planar surface. The paleosol is characterized by a pronounced enrichment of iron oxides which gives the sediment a reddish brown to in places an almost black coloration. The paleosol is overlain by a lighter colored, medium grained, moderately sorted, near symmetrical to coarse skewed, mesokurtic to leptokurtic, quartz sand. This sand appears to be eolian in origin. This sand probably corresponds to the "Castalia sand" of Wells (1944). The Castalia sand has a weak soil zone developed upon it. Throughout the remainder of this report the informal stratigraphic names of Wells (1944) will be applied.

Sometime after deposition of the shell debris and prior to the deposition of the Castalia sand the Cape Fear shell debris were subjected to two stages of diagenesis. During the first stage the Cape Fear coquina was partially lithified under a fresh water vadose to phreatic environment. The pendent cements of the unlithified shell hash are indicative of a vadose regime

A sample of dog tooth calcite crystals from the Ft. Fisher Historic Site area (Figure 1) was also made in conjunction with this project. The outcrop from which the dated material was obtained is found in the intertidal zone of the modern beach immediately east of the parking lot for the Visitors Center. Here the coquina is more sandy than at Snows Cut and has more readily visible sedimentary structures including trough crossbeds. Like at Snows Cut, it is overlain by the moderate to poorly sorted Kure sand, a paleosol, and a fine well sorted sand. The crystals were collected and inspected in an identical manor to that of the Snows Cut crystal sample. However, the material was not as fresh as the material from Snows Cut. A number of the crystals had to be rejected due to the presence of modern algae on them. The sample submitted for radiocarbon analysis consisted of 50 grams of material (approximately 940 crystals) taken from three shells.

A *Mercenaria* sp. shell from an abandoned and now filled borrow pit known as the Burnett Pit (Figure 1) was collected by Victor Zullo and radiocarbon dated circa 1978. Laboratory number and sample size are no longer available. Material near the site of the pit would suggest that the sample came from a lithology identical to the lower or cemented layer at Snow's Cut.

The specimen of *Crassostrea virginica* was obtained by William Cleary from a coquina layer encountered at a depth of 4.5 meters in a vibracore taken in the salt marsh at the northern end of Carolina Beach, NC approximately 2 kilometers southeast of the Snows Cut site (Figure 1). The bed from which it was collected is correlative to the Cape Fear Coquina at Snows Cut and lithologically similar to the shell hash lithofacies. The 4.5 meters of material which was overlying the dated sample was typical barrier island-salt marsh sediments which in the area have been accumulating for four to five thousand years.

EVALUATION OF CONTAMINATION

The Snows Cut samples present apparent radiocarbon ages which are much younger than what is indicated by other methods of dating. The most simplistic explanation of this would be that the dated material was contaminated by a small amount of younger carbon. If, for example, a shell with an absolute age of 124 ka B.P. is contaminated by just one percent modern (<100 years in age) carbon the resultant apparent radiocarbon age would be about 37 ka B.P. (Figure 10). If the shells from Snows Cut do indeed date from the time of the last generally accepted marine high stand or approximately 124 ka (absolute) B.P. then the *Nassarius obsoleta* sample would have a minimum modern carbon contamination of 5%, the *Crassostrea virginica* sample 2.5%, and the *Donax variabilis* sample 2%.

Contamination of marine shells can occur by the (a) growth of rootlets on, around, and within the shells, (b) deposition of cements on the exterior surfaces of the shells, (c) infilling of borings within the shells by younger material, or (d) replacement of biosecreted aragonite by diagenetic calcite. At the time of collection of the samples, rootlets were totally absent from the deposit which was situated well below the modern soil zone. Rootlets are now, several years after the collection of the samples for radiometric dating, present at the outcrop site. These are dark gray to dark brown and show very well against the very light buff to white color of the shells. There is virtually no chance that rootlets contaminated the dated material given the fresh nature of the collected material and the ease of recognition of rootlets. The shells were all collected from that part of the section where there were minor amounts of pendent cement deposited on the underside of some shells. The meniscus, bladed, and blocky calcite cements all occurred at a stratigraphic horizon below that from which the radiocarbon samples were obtained. As noted previously the *Donax variabilis* shells were field checked for visual evidence of pendent or vadose cements and those with such were discarded. For a typical shell with pendent cement and which would have been discarded, the cement represented no more than 5% of the $CaCO_3$ volume of the shell. As at most only 1% of the *Donax variabilis* shells in the strata exhibited the pendent cements, the level of contamination from this source could therefore not have exceeded 0.05%, even if the shells were not hand picked. However, such shells with pendant cement were discarded therefore the level of contamination from this possible source must be considerably less. A similar situation existed for the *Nassarius obsoleta* shells. However with these, visual inspection was more tenuous due to the structure of the gastropod shell and therefore the non shell $CaCO_3$ could have been as high as 0.05% of the sample volume. The *Crassostrea virginica* sample, when collected, had a significant deposit of pendent calcite cement which represented no more than 1% of the volume of the sample. However, prior to submitting the sample for radiocarbon dating, the sample was extensively etched with HCl such that 40% of the sample mass was removed. This included all the pendent cement along with the entire surface layer of the oyster shells. Borings and microborings in the shells were common throughout the section. From modal analysis of the samples (Table 3) borings would have represented at most 5% of the volume of a typical carbonate shell. Thin section examination of the splits from the radiocarbon samples and other material from the horizon from which the radiocarbon samples were taken revealed that within the resolution of the petrographic microscope these borings, in all cases, lacked any infilling material; either sediment or cement (Figure 5B). Even within the zone where there was significant meniscus cements the borings still remained open even though only a few microns away occurred calcite cement material. Only borings from the lower horizon with the blocky and bladed calcite cement exhibited filling by calcite. Assuming the resolving power of the microscope would allow one to see at least a 5% filling of these borings the maximum conceivable volume of calcite contamination would have only been 0.25%. The diagenetic replacement of aragonite by calcite is a common feature of ancient marine sediments. The replacement process is analogous to that of dolomitization of calcite as described by Dockal (1989). During the replacement process some isotopic fractionation is to be expected. During calcitization of aragonite isotope fractionation results in the enrichment of the secondary calcite with the heavier isotopes of carbon. However, the only material at Snows Cut with calcification of aragonite was found from the cemented material below the horizon from which the dated shell material were taken.

Isotopic exchange between the pore waters and the surface of the shells and crystals might be possible, but there is no evidence to support such. Had such happened then the shells with the greatest surface area relative to mass, *Donax variabilis*, would have been affected the most and would have had the youngest apparent radiocarbon age; the *Dona*x shell sample gave the oldest age. Furthermore, such effects should have been eliminated or reduced by the pretreatment at Beta Analytic.

DISCUSSION

Contamination of these samples, from the foregoing discussion, seems unlikely and in the *worst possible case* could not have exceeded 0.25% of the volume of the sample. Such a level of contamination, if assumed to be entirely modern (<100 years), would give an apparent radiocarbon age of 47 ka B.P. assuming the shell material was 124 ka in absolute age (Figure 10). If, on the other hand, the absolute age of the shells is something like 124 ka and the contaminating carbon was again modern the indicated contamination level would have to be 2% to 5%. However, the most likely contaminant, the calcite cements, could not have been entirely modern and for the most part had to form at a much earlier time. The cross-cutting relationship of the dissolution fingers associated with the formation of the Kure sand and the cemented portion of the Cape Fear coquina indicates that the calcite cements formed prior to this dissolution feature. Furthermore the planar nature of the paleosol at the top of the Kure sand and below the Castalia sand indicates that the fingers and consequently the cements predate that soil forming process. Formation of the cements probably was completed thousands of years ago. Furthermore, the calcite cements themselves did not happen in an instantaneous manor but also developed over a period of time measured in hundreds if not thousands of years. This requires us to talk in terms of a *mean* age of cement formation which reflects the time the process began, its duration and rate, and the time of its termination. If the mean age of the contaminating calcite cementation is 10 ka B.P. then the requisite level of contamination would be in the order of 8% to 18% and if the mean age is 20 ka B.P. that becomes 25% to 60% (Figure 10). Contamination of shell samples to this degree would be *blatantly obvious*. As noted previously there is no petrographic evidence for contamination of the Snows Cut material beyond a maximum of 0.3% and that only applies to the sample of *Nassarius obsoletus*.

The dog tooth calcite found within the shelter porosity of *Mercenaria* and *Busycon* shells is directly related to the blocky and bladed calcite cements and the calcitized aragonite bioclasts found just below the horizon from which the dated shells were taken. The cements within the Snows Cut section form a suite of diagenetic features that formed simultaneously at or near the position of the paleowater table and predate development of the Kure sand. These cements represent the most obvious and most probable contamination source. The dog tooth calcite should be considered the *most grossly contaminated* material within the section and furthermore, in comparison to the level of contamination of the dated shells, this sample of calcite crystals should be considered 100% by volume contaminated material. Keep in mind the shell samples were considered significantly less than one percent by volume contaminated with diagenetic calcite cement. The apparent radiocarbon age on the dog tooth calcite is 26,380+/-500 yr. B.P. There are two possible sources of carbon for formation of this calcite: (1) that derived from dissolution of the shells within the bed and (2) that brought into the strata from the atmosphere via downward moving meteoric water. The carbonate of the calcite crystals will be a mixture of carbon from both sources. As modeled by Garrels and Christ (Case 5, p.86-88, 1965) and Dockal (1992) meteoric water is brought to equilibrium with atmospheric carbon dioxide with a resultant total carbonate concentration ($[H_2CO_3] + [HCO_3^-] + [CO_3^=]$) of $10^{-4.8}$ moles per liter. This meteoric water travels downward and interacts within the shell beds below the water table and causes dissolution of the aragonitic shells, calcification of aragonite, and precipitation of calcite cements to a degree such that the total carbonate concentration must now reach $10^{-3.8}$moles per liter. The dissolved carbonate is the sum of that brought into the shell beds via meteoric water and that derived from aragonite dissolution. The percentage of carbon of meteoric origin that goes to the precipitating calcite should therefore be approximately ($10^{-4.8}/10^{-3.8}$) or 10% whereas that derived from dissolution of the shells, especially the aragonitic shells, would be 90%. If such is the case and if the shells are approximately 124 ka in absolute age then the mean apparent age of the cementation process would be about 2 ka. This, however, also requires that the shells now also be 100% diagenetic calcite for which there is no petrographic evidence. Increasing the partial pressure of CO_2 to accommodate soil gas does not change this situation much; it only pushes the mean age of cementation back a few thousand years and still requires the shells to be 100% diagenetic calcite. Furthermore the development of the paleosol above the Kure sand, the accumulation of the Castalia sand, and the development of a soil at the top of the Castalia sand all would have had to have happened in an unreasonably short period of time. If, on the other hand, one takes the weighted mean apparent

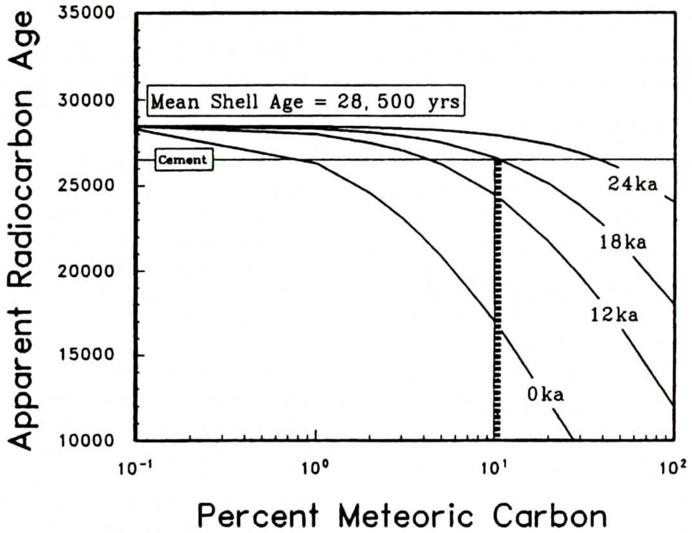

Figure 11. The apparent radiocarbon age of post depositional calcite cement (vertical scale) is a function of the ages of the sources of the carbon for the cement; meteoric water and shell debris. Shell source material has an living age of 28,500 years. Curves represent mean times of cement growth. Horizontal scale gives the percent carbon derived from meteoric sources. Horizontal line labeled "Cement" represents the age of the cements in the coquina at Snows Cut (Beta-37434) which if they consisted of 10% meteoric carbon would have to have a mean age of cement growth of approximately 18 ka B.P. Actual growth probably was initiated prior to this time and was completed later.

radiocarbon age of the shells (28,500 yr. B.P.) to be a first approximation of their actual *apparent radiocarbon age* then the apparent age of the calcite crystals would imply that the calcite cementation process has a mean age of 18 ka B.P. (Figure 11). This would result in a second approximation of shell radiocarbon age to be about 30 ka B.P. for a diagenetic calcite contamination level of 1% by volume; which is more than is indicated by petrographic evidence. The 30 ka B.P. age is the oldest conceivable *apparent radiocarbon age* that uncontaminated shells from this deposit could have with the exception of those fossils (remanie death assemblage) within the bed that were reworked from older units.

It is possible but unlikely that these young apparent radiocarbon ages are the result of a laboratory contamination problem. The ages reported here were determined by Beta Analytic Inc. which has a solid reputation of quality work. Furthermore, the samples were run at two different times: Beta-37431, 37432, 37433, and 37434 reported May 31, 1990; while Beta-38473 was reported July 30, 1990. It is unlikely that a systematic problem would have persisted at the lab for two months and it is even more unlikely that a random problem would have appeared to affect both sample sets. The other radiocarbon ages from the Cape Fear coquina (Table 1) further shed doubt upon this.

CONCLUSIONS

The radiocarbon dated shell samples from Snows Cut have no significant levels of contamination by younger carbon. The highest conceivable level of contamination is 0.3% by volume which includes both contamination from the pendent cements and the filling of part of the borings by calcite; neither of these however, are supported by petrographic evidence. If the samples were indeed contaminated by 0.3% modern (< 100 years) carbon and the shells actually dated from 124,000 absolute years B.P. then they should have presented an apparent radiocarbon date of 47,000 yr. B.P. The contaminating material itself, the calcite cements, has a mean apparent radiocarbon age of 26 ka B.P. The oldest that the shells could possibly be is 30,000 *radiocarbon years before present* and the diagenetic process that produced the calcite cements has an approximate mean age of 18 ka B.P. There is no way the shells at Snows Cut which were contemporaneous with deposition of the Cape Fear coquina could date from 124,000 radiocarbon years B.P. If the Cape Fear coquina was deposited 124,000 absolute years B. P. or even 60,000 absolute years B. P. as suggested by Prosser (1993) then something other than contamination has affected the radiocarbon age determination of this material.

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THE CARPENTER FORK BED, A NEW — AND OLDER — BLACK-SHALE UNIT AT THE BASE OF THE NEW ALBANY SHALE IN CENTRAL KENTUCKY: CHARACTERIZATION AND SIGNIFICANCE

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ABSTRACT

Black shales previously interpreted to be Late Devonian cave-fill or slide deposits are shown to be much older Middle Devonian black shales only preserved locally in Middle Devonian grabens and structural lows in central Kentucky. This newly recognized — and older — black-shale unit occurs at the base of the New Albany Shale and is named the Carpenter Fork Bed of the Portwood Member of the New Albany Shale after its only known exposure on Carpenter Fork in Boyle County, central Kentucky; two other occurrences are known from core holes in east-central Kentucky. Based on stratigraphic position and conodont biostratigraphy, the unit is Middle Devonian (Givetian: probably Middle to Upper *P. varcus* Zone) in age and occurs at a position represented by an unconformity atop the Middle Devonian Boyle Dolostone and its equivalents elsewhere on the outcrop belt. Based on its presence as isolated clasts in the overlying Duffin Bed of the Portwood Member, the former distribution of the unit was probably much more widespread — perhaps occurring throughout western parts of the Rome trough. Carpenter Fork black shales apparently represent an episode of subsidence or sea-level rise coincident with inception of the third tectophase of the Acadian orogeny. Deposition, however, was soon interrupted by reactivation of several fault zones in central Kentucky, perhaps in response to bulge migration accompanying start of the tectophase. As a result, much of central Kentucky was uplifted and tilted, and the Carpenter Fork Bed was largely eroded from the top of the Boyle, except in a few structural lows like the Carpenter Fork graben where a nearly complete record of Middle to early Late Devonian deposition is preserved.

INTRODUCTION

The stratigraphic interval encompassing the New Albany-Chattanooga-Ohio shale is probably one of the most distinctive and best known stratigraphic intervals throughout the east-central United States. Although each of the above units largely represents the same, central Late Devonian interval of black-shale deposition, the age of each varies somewhat at its top and bottom (Figure 1). The Ohio Shale in northeastern and parts of eastern Kentucky (Figure 2) is wholly Late Devonian (Famennian) in age; the black, Lower Mississippian (Tournaisian, Kinderhookian) Sunbury Shale is separated from it in this area by the gray shales and siltstones of the Upper Devonian Bedford and Berea formations (Ettensohn and others, 1979, 1988; Pashin and Ettensohn, 1995) (Figure 1).

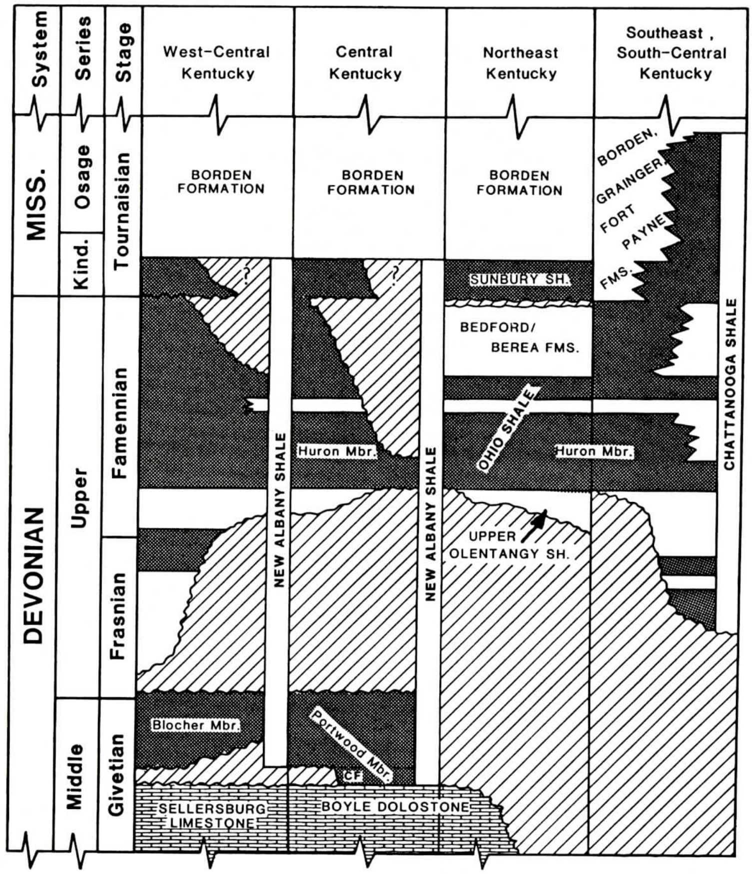

Figure 1. Nomenclature and approximate regional correlations between Devonian-Mississippian black-shale units in Kentucky. Dark pattern represents section with predominantly black shales; white areas in columns represent section with predominantly gray or green shales, siltstones or sandstones; brick pattern, carbonates; diagonal rule, missing section; CF, Carpenter Fork Bed (Mississippian portion modified from Elam, 1981; Ettensohn and others, 1988; and Lierman and others, 1992).

In southeastern and south-central Kentucky (Figure 1), however, the Bedford-Berea sequence becomes thinner and grades into black shales so that Lower Mississippian Sunbury black-shale equivalents conformably overlie black-shale equivalents of the Bedford and Berea which, in turn, conformably overlie Upper Devonian, Ohio Shale equivalents. The entire interval then becomes an Upper Devonian-Lower Mississippian unit called the Chattanooga Shale (Elam, 1981; Ettensohn and Elam, 1985) (Figures 1 and 2). In south-central parts of Kentucky, however, Mississippian black shales are absent from the Chattanooga due to nondeposition or facies changes into gray shales in lower parts of the Fort Payne or Borden formations (Ettensohn and others, 1989) (Figure 1).

The major difference between the New Albany Shale of western and central Kentucky and the Chattanooga Shale of south-central, west-central, and southeastern Kentucky is that in western and central Kentucky, Middle Devonian (Givetian; Fingerlakesian) black shales are present at the base of the sequence (Lineback, 1968, 1970; Ettensohn and others, 1988, 1989) (Figure 1). In these cases, the Middle Devonian-through-Lower Mississippian black-shale sequence is called the New Albany Shale (Figures 1 and 2). In western and west-central Kentucky, the Middle Devonian part of the New Albany is included in the Blocher

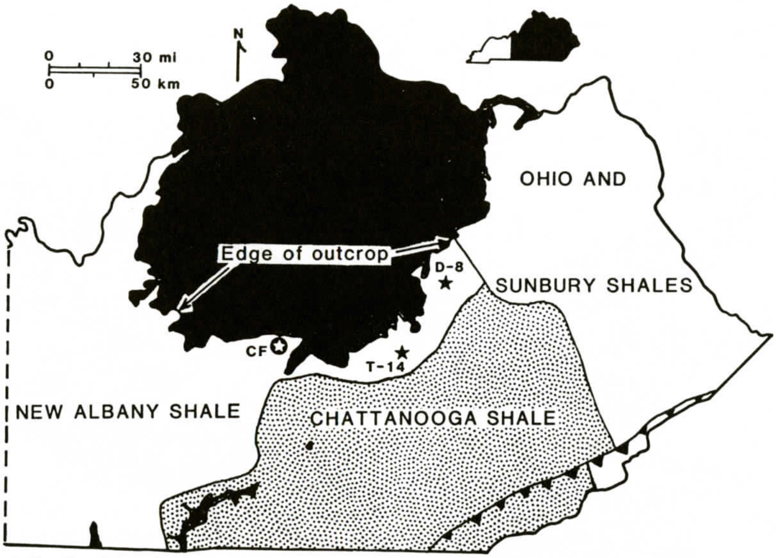

Figure 2. Areal distribution of Devonian-Mississippian black-shale units in Kentucky. Black pattern represents exposure of pre-Devonian rocks. Encircled star is the Carpenter Fork (CF) section; black stars locate the two cores containing the Carpenter Fork Bed (see Appendices).

Member (Givetian; Fingerlakesian), whereas in central and east-central Kentucky, the Middle Devonian portion was included in the Portwood (or Trousdale) Member (late Givetian) by Campbell (1946) (Figure 1). Although the name Portwood has gained some acceptance in the stratigraphic literature, the Trousdale designation was never widely accepted, and black shales assigned by Campbell (1946) to the Trousdale are generally included in the Portwood. The distribution of the Portwood is restricted to structural lows along the Kentucky River Fault System (Ettensohn and others, 1989) (Figure 3), and the unit unconformably overlies the Boyle Dolostone (Givetian; Tioughniogan and early Taghanic) (Figure 1), commonly with a subtle angular unconformity.

The Portwood and its Blocher equivalents were previously considered to be the oldest black shales in the New Albany; however, basal, organic-rich dolomitic breccias of the Duffin Bed of the Portwood Member (Campbell, 1946) are known to contain black-shale clasts suggesting the former presence of an even older, underlying black shale. In this study, we describe an exposure along Carpenter Fork in Boyle County, central Kentucky (Figure 4), containing this older black-shale unit, which is gradational between the Boyle and the Portwood. Although the unit is known only from a single locality and two cores, the outcrop occurrence in a well-defined Devonian graben and the fact that it was deposited during a time which is represented elsewhere in the area by a hiatus (Boyle-Portwood unconformty) indicate that the unit and its circumstances of occurrence could be significant in elucidating aspects of the structural, tectonic, and depositional history of the area that were not previously known. Moreover, because the deposition of the Devonian-Mississippian black shales was closely related to the Acadian orogeny (Ettensohn, 1985, 1987; Ettensohn and others, 1988), and because this shale unit may shed light on the effects of the orogeny in a cratonic area, we believe that it is important to name and characterize the unit in enough detail that such interpretations are possible.

LOCATION AND REGIONAL STRUCTURE

In central Kentucky, the Devonian outcrop

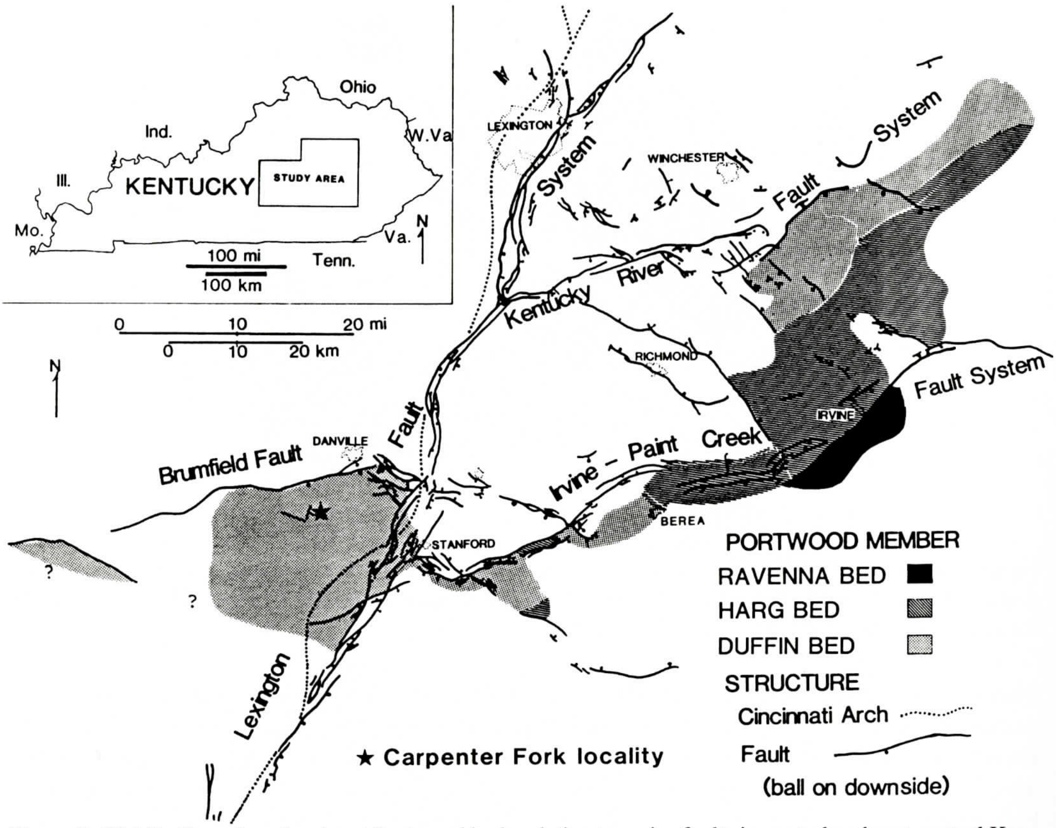

Figure 3. Distribution of predominant Portwood beds relative to major faults in central and east-central Kentucky (adapted from Ettensohn and others, 1991).

exhibits predominantly carbonates and organic-rich shales. These shales form an arcuate belt in central Kentucky (Figure 4), which generally coincides with the physiographic region known as "The Knobs" (Figure 4). The studied shale unit is known from one five-meter thick exposure in the south-central part of the outcrop belt in southern Boyle County, along the Carpenter Fork of the North Rolling Fork River, one-kilometer north of the Casey County line (Carter Coordinates: 2650 ft FSL x 100 ft FWL, 14-M-56) (Figure 4). Thinner sections less than two-meters thick are known from cores in Powell County (Carter Coordinates: 2775 ft FSL x 400 ft FEL, 6-P-68) and Jackson County (Carter Coordinates: 2000 ft FNL x 1000 ft FWL, 11-M-64) (Figure 2).

The arcuate pattern of the outcrop belt is related to uplift and erosion on the Jessamine dome, a structural high on the axis of the Cincinnati arch in north-central Kentucky. The entire central Kentucky region has also been subjected to recurrent faulting on the Lexington, Kentucky River, Irvine-Paint Creek, Brumfield and other, unnamed fault systems (Figure 4). These fault systems were active at intervals throughout the Paleozoic, and may have repeatedly changed their sense of movement, as is evident on some faults in the immediate area of Carpenter Fork (Moore, 1978).

DEVONIAN STRATIGRAPHY OF CENTRAL KENTUCKY

The Devonian stratigraphy of central Kentucky is characterized by Middle and Upper Devonian strata which onlap an uneven erosion surface on Ordovician and Silurian rocks. The onlap of Devonian strata across pre-Devonian

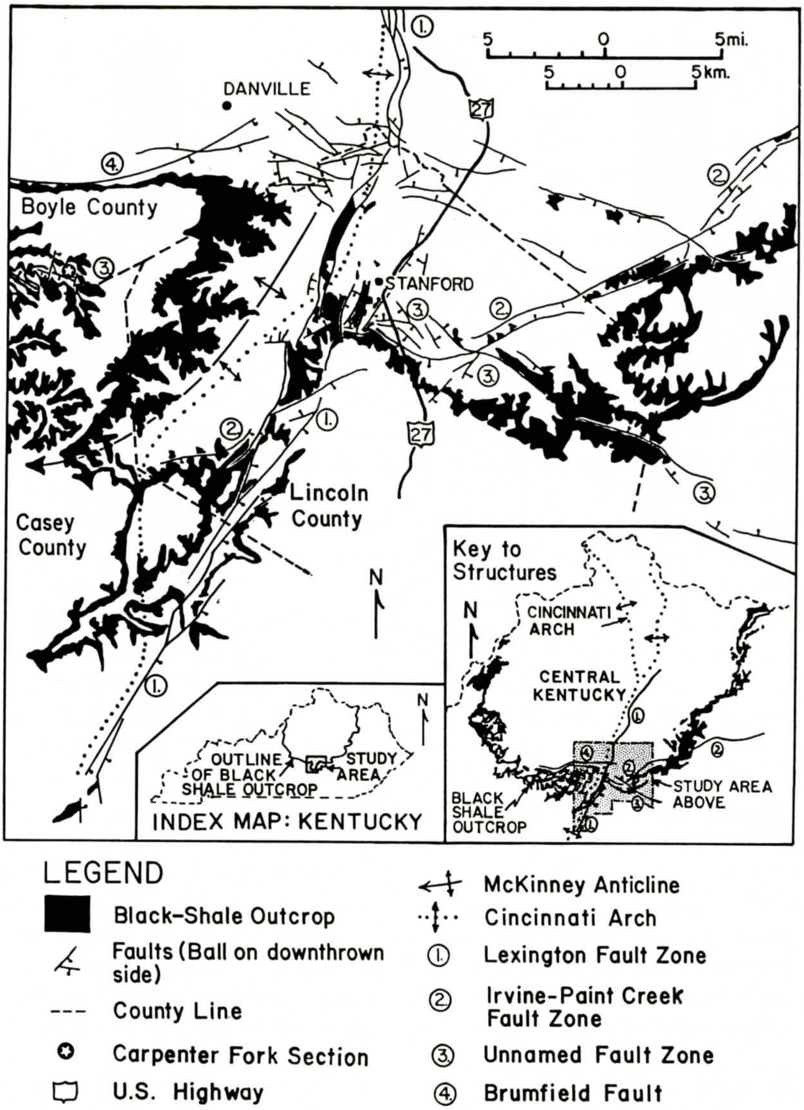

Figure 4. Location map showing the distribution of Devonian-Mississippian black shales, regional structures, and location of the Carpenter Fork section (encircled star) in central Kentucky; the arcuate black-shale outcrop belt largely corresponds to the Knobs physiographic province (adapted from Ettensohn and Bayan, 1990).

units is disconformable on the outer margin of the Jessamine dome but becomes angularly unconformable toward the interior outcrop margin. Although the Devonian sedimentary sequence generally reflects a period of transgression, there are indications of multiple regressive, erosional or hiatal events marked by sandy or phosphorite- and bone-rich, interbedded lag horizons (Conkin and others, 1976; Lenhart, 1985). The typical stratigraphic column for central Kentucky is shown in Figure 5A.

The lowermost unit, the Middle Devonian (Eifelian-Givetian) Boyle Dolostone (see Woodrow and others, 1988) has been interpreted by recent authors to be a shallow-subtidal deposit (*e.g.*, Lenhart, 1985). The Boyle is primarily a fine- to medium-grained dolarenite

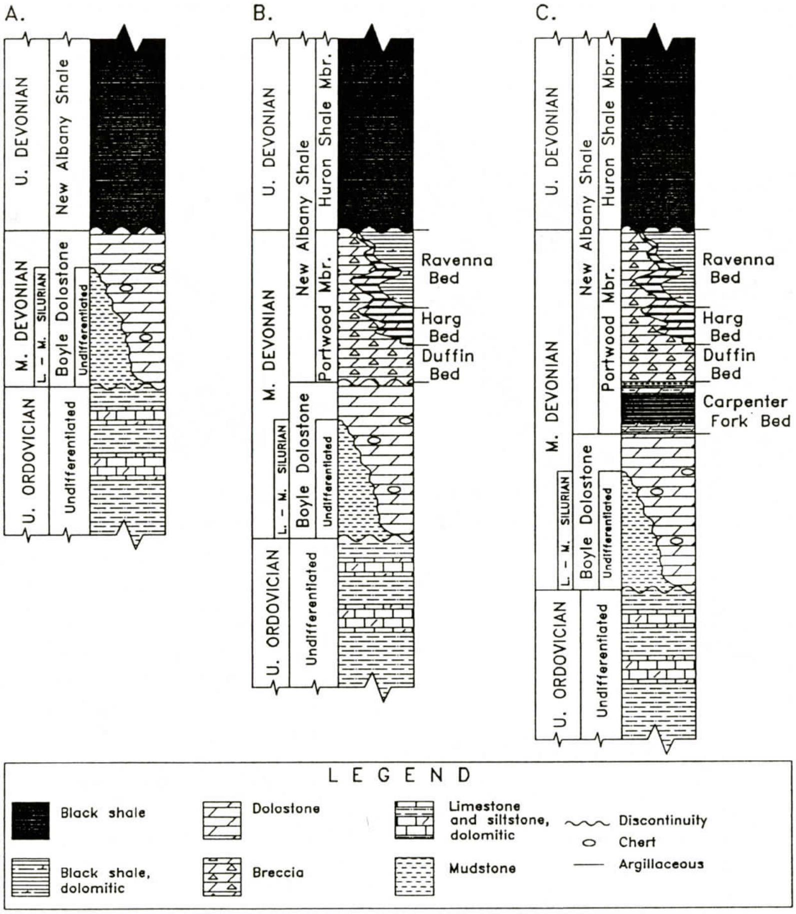

Figure 5. Three examples of Middle Devonian sections from central Kentucky: A.) Upper Devonian black shales disconformably overlying Middle Devonian Boyle carbonates; B.) Middle Devonian Portwood beds disconformably overlying Boyle carbonates and disconformably overlain by Upper Devonian black shales; C.) Carpenter Fork black shales gradational between Boyle carbonates and other overlying Portwood beds.

with variable amounts of chert, sand, and silt. The Boyle was deposited on a surface with several meters of relief; for example, the Boyle laps onto the flanks, but not the tops, of "fossil" Silurian outliers in east-central Kentucky (McFarlan and White, 1952; Moore, 1978). The Boyle, in turn, is typically overlain unconformably by the Upper Devonian (late Frasnian to middle Famennian) upper Olentangy Shale equivalent or Huron Shale Member of the New Albany Shale (Figures 1 and 5) or by one of three different beds of the Portwood Member of the New Albany Shale (*e.g.*, Barnett and Ettensohn, 1992) (Figure 5B). The Portwood Member, as reinterpreted by Ettensohn and others (1988, 1989), is of Middle Devonian (late Givetian) age (Figure 1) and consists of three largely coeval facies or beds (Figures 3 and 5). The Duffin Bed is predominantly an organic-rich dolomitic breccia with clasts of dolostone and chert, as well as rare siltstone and black-shale clasts. The lithology and fossil content of the dolostone, chert, and siltstone clasts indicate that they were derived from the underlying Boyle, but the origin of the dark-shale clasts has been problematic because, previously, no black shales were known to occur below the Duffin. Clast size is typically less than two centimeters, but the lowermost Duffin commonly contains clasts in the two- to sixcentimeter range and, less frequently, up to boulder size. Locally, these boulders attain sizes in excess of two meters in their long dimension. The lithoclasts are matrix supported, generally angular, and lighter in color than the supporting matrix. Although the dolomitic matrix is generally dark and organic-rich, locally the matrix is less organic-rich and the clast-matrix distinction is obscured. Chert clasts are generally angular, may show weathering rinds, and tend to be less abundant than dolostone clasts, although the lower Duffin locally consists almost entirely of densely packed chert clasts. Shale and siltstone clasts are angular to rounded and are generally oriented subhorizontally. Some of the flatter clasts clearly show signs of imbrication and preferential orientation.

Two other facies of the Portwood Member, the Harg and Ravenna beds, consist of variable amounts of gray to black shale, dolomitic mudstone, and argillaceous to arenaceous limestones and dolostones. In Boyle County and nearby areas, the Duffin Bed is typically conformably overlain by Harg and Ravenna beds. These, in turn, are overlain paraconformably by the Upper Devonian upper Olentangy Shale equivalent or Huron Shale Member of the New Albany Shale (Figures 1 and 5).

The unconformity between the Portwood and the upper Olentangy or Huron is marked by a lag zone, generally thin and obscure, typically consisting of rounded and frosted quartzose sand, phosphorite grains, pyrite, and abundant conodont elements. Artificial gamma-ray logs prepared with a hand-held scintillometer (Ettensohn and others, 1979) reveal that the unconformity is marked by a pronounced change from low-radioactivity Middle Devonian black shales below to high-radioactivity Upper Devonian black shales above. A radiation "kick" can generally be detected just above the lag horizon.

LOCAL STRUCTURE

The Carpenter Fork section is exposed in a creek-cut cliff about 50-m wide, trending roughly east-west. The exposure is unique in that it contains a relatively thick (5.4 m), section of fissile black shale sandwiched between the Boyle below and the Duffin above in what appears to be a Devonian graben (Figure 6). Above the Duffin are beds of black shale in the usual stratigraphic position (Figures 5C and 6).

Regional structures which may have had local influence on deposition include the Cincinnati arch, the axis of which trends north-northeast to south-southwest about 12-km east of the study area; the Lexington fault zone, a series of faults near the crest of the arch (Figure 4, no. 1); the east-west trending Brumfield fault, down-thrown to the south and located about 6.6-km north of the Carpenter Fork section (Figure 4, no. 4); and an unnamed northwest-southeast-trending fault zone (Figure 4, no. 3) that passes through the Carpenter Fork area. The study area is near the southern margin of what Lenhart (1985) termed the Fork-land trough, a graben-like feature whose indistinct southern boundary was suggested by the thinning or absence of lower members of the Boyle Dolostone south of the study area.

In addition to the broader structural features discussed above, many smaller scale faults and gentle folds are common in the local area (Figure 7). Offset on these faults is typically on the order of a few meters or less. The Carpenter Fork creek-side exposure appears to be a structurally truncated graben as indicated by displacement of strata, apparent drag folding on beds close to bounding faults, and the mapped distribution of units (Figure 7). As seen in Figure 7, the graben is part of a complex of faults which have produced a displacement of 13 m or more between the Boyle at the base of the graben and the Boyle on the hillside east of the graben; on the western end of the graben, the Boyle has been offset by at least 2.5 m. The Boyle Dolostone, exposed on the uplifted margins of the graben, consists of light-gray, fine-grained dolarenite with chert nodules. Within the graben, disturbed beds of similar lithology are found at the base of the section close to the bounding faults, but they dip below creek level

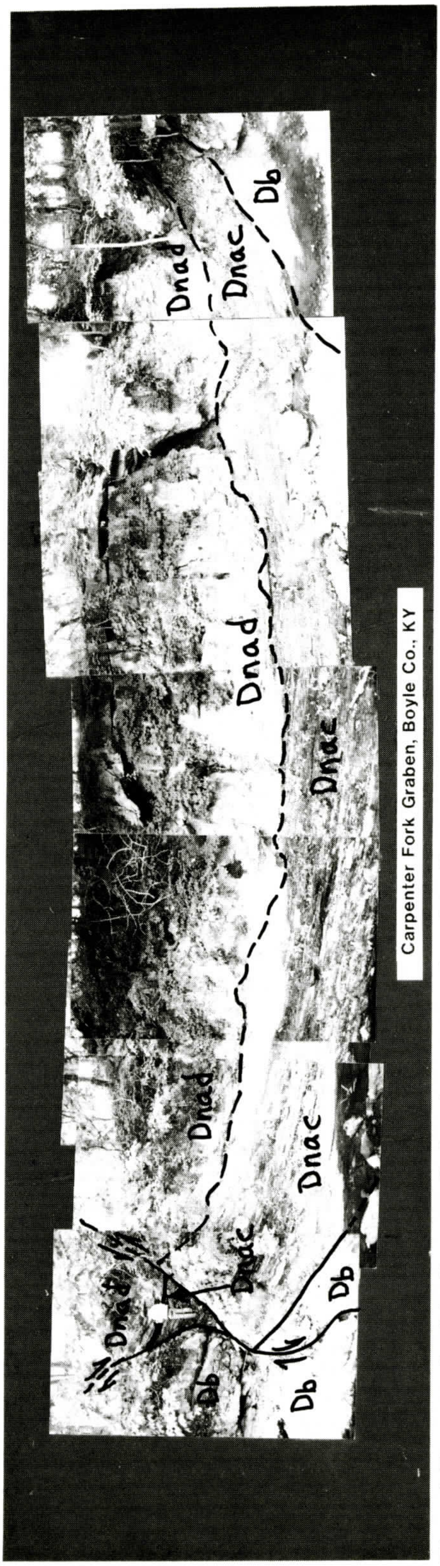

Figure 6. Photomosaic of the Carpenter Fork graben, type section of the Carpenter Fork Bed; unit designations and structure superimposed on photo. Southeastern edge of graben not shown. Db, Boyle Dolostone; Dnad, Duffin breccia; Dnac, Carpenter Fork Bed. (Photos by C. Mellon).

toward the interior of the graben (Figure 6).

Although beds outside the graben are roughly horizontal, within the graben all beds are upturned against the faults (Figure 6). What is unusual about the Carpenter Fork graben is the presence of a black-shale unit between the Boyle and the Duffin Bed of the Portwood Member (Figure 5C). This black shale is gradational with the underlying Boyle and has not been reported previously as a new unit. The nature and origin of this unit, which we herein designate as the Carpenter Fork Bed of the Portwood Member, is the subject of the remainder of this paper.

THE CARPENTER FORK SECTION

We herein designate the section in the Carpenter Fork graben as the type section of the Carpenter Fork Bed of the Portwood Member of the New Albany Shale; the section is formally described in Appendix A and illustrated schematically in Figure 8. The unit is defined as a black shale gradational with the underlying Boyle Dolostone and occurring below either the Duffin or Harg beds of the Portwood Member (Figures 5C, 6, and 8). The lowermost beds exposed in the graben (unit 1, Appendix A) are of typical Boyle lithology, consisting of fine-grained, locally cherty dolarenite. The upper portion of the Boyle (unit 2) becomes increasingly argillaceous and sandy, and contains sand- and granule-sized phosphorite debris. Overall, the unit coarsens upward and becomes more thinly bedded. The unit is locally micro-cross-laminated with organic-rich flaser beds and is extensively burrowed. The contact between the Boyle and the Carpenter Fork Bed is gradational. The sandy dolostone of the Carpenter Fork (unit 3) is differentiated from the subjacent sandy dolostone of the Boyle on the basis of organic matter: the Carpenter Fork dolostone is darker than that in the subjacent Boyle, although the contact is gradational. The basal Carpenter Fork Bed is also more quartz-rich and less phosphatic than subjacent Boyle units and is characterized by thin dark-shale partings, hummocky cross-beds, and scours. It

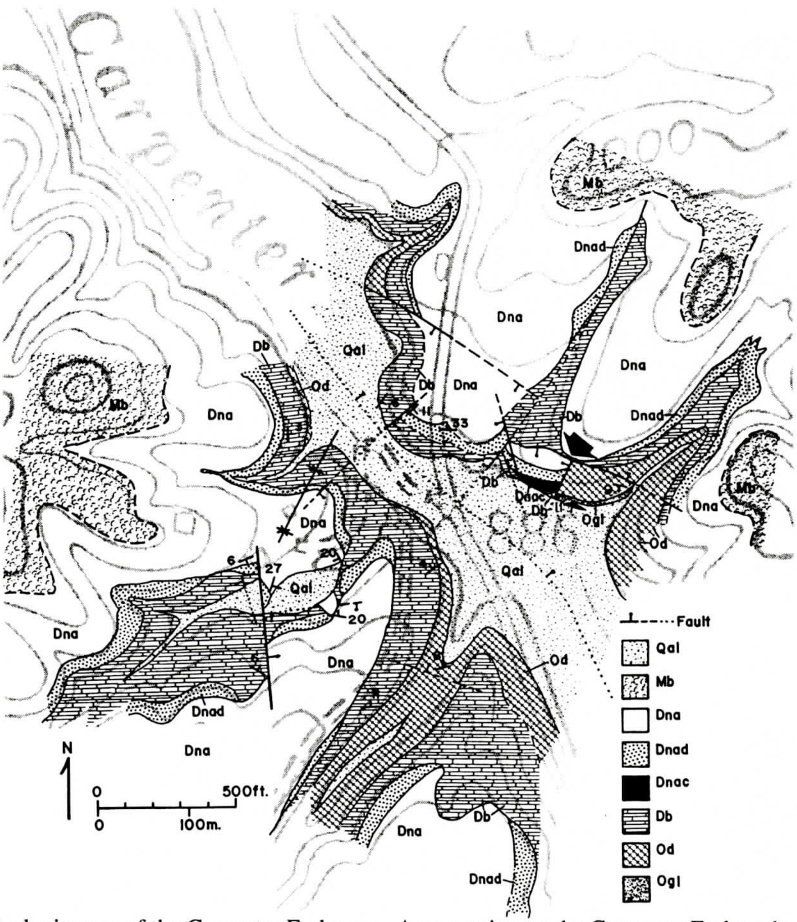

Figure 7. Geologic map of the Carpenter Fork area. Arrow points to the Carpenter Fork graben. Note that most of the Duffin breccias (Dnad) are present on the downdropped sides of faults. Ogl, Grant Lake Fm.; Od, Drakes Fm.; Db, Boyle Dolostone; Dnac, Carpenter Fork Bed; Dnad, Duffin Bed; Dna, overlying Middle and Upper Devonian black New Albany shales; Mb, Borden Fm.; Qal, alluvium.

is locally burrowed and contains spiriferid brachiopods. The lower, sandy unit of the Carpenter Fork grades upward through a few centimeters into fissile, silty, grayish-black shales and minor, brownish-black, interbedded dolosiltites which comprise the bulk of the member (3.6 m, unit 4). The shale is organic-rich and pyritic throughout, although lingulate brachiopods and burrows are common, especially in conjunction with dolosiltite beds. Dolosiltite beds become more abundant in the upper two meters of the unit, and about one-half meter from the top of unit 4, pyritic, dolomitic siltstone beds become especially abundant. These beds are sparsely glauconitic and exhibit sparse micro-cross-laminae, wavy bedding, and burrows. Sandy, organic-rich dolstone beds, interbedded and interlaminated with dark shales similar to those below, occur near the top of the Carpenter Fork Bed (unit 5). The dolostones are glauconitic and pyritic and exhibit scours, ripples and micro-cross-laminae, as well as wavy, flaser and lenticular beds. Burrows, including *Cruziana* and escape burrows, are very extensive (Figure 9). The thickness of the unit varies from 0 to 0.7 m; the beds thin or have been truncated toward the margins of the graben.

Unit 6 is also of variable thickness, ranging from 0 to about 0.3 m. Lithologically, it is very similar to the black shales of unit 4, except that locally it contains allochthonous clasts of chert.

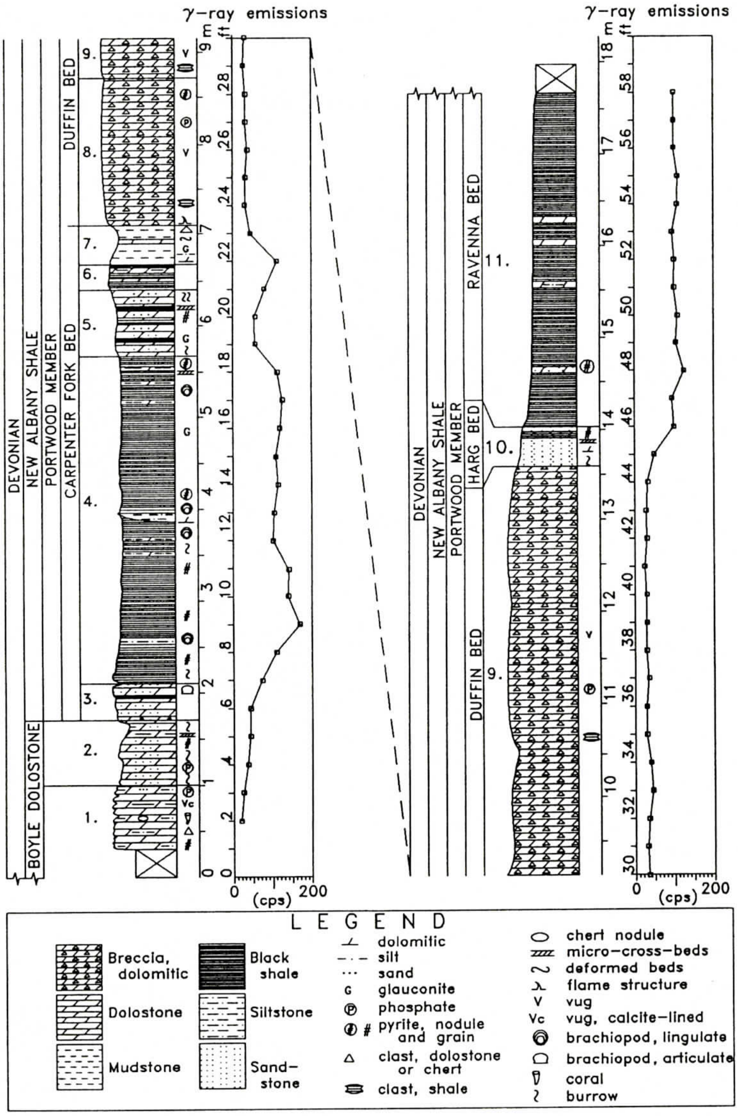

Figure 8. Schematic stratigraphic column and radioactivity profile from the type section of the Carpenter Fork Bed. Numbered units are described in Appendix A.

The uppermost unit of the Carpenter Fork Bed (unit 7) is an organic-rich, dolomitic mudstone which has been subjected to extensive soft-sediment deformation. Within the mudstone, much of the deformation was apparently caused when clasts of chert, up to 0.7 m in long dimension, were dumped onto the muds (Figure 10). The mudstone was injected up into the spaces between clasts of chert, and locally into the overlying dolomitic breccia due to loading. In a few places the mudstones have been folded and overturned (Figure 11).

Above the Carpenter Fork Bed are dolomitic breccias and dark shales of the Duffin, Harg, and Ravenna beds of the Portwood Member of the New Albany (Figure 8). The lower breccia unit (unit 8) has atypically large clasts of dolostone and chert. Clasts of Boyle Dolostone range up to two meters in long dimension. The unit also contains clasts of black shale. The

Figure 9. Intense bioturbation in the sandy, organic-rich dolarenites from Unit 5 of the Carpenter Fork Bed (See Figure 8 and Appendix A.).

Figure 10. Dark organic-rich mudstones from Unit 7 of the Carpenter Fork Bed injected upward between light-colored, allochthonous, cobble-size Boyle chert clasts at the base of the Duffin (See Figure 8 and Appendix A.). Contact between injected mudstone and chert clasts outlined with a dashed line. Flame structure just above hammer point.

matrix of the breccia is an organic-rich, silty dolostone. The overlying unit 9 includes from one to five meters of more typical Duffin breccia, containing fewer chert clasts and clasts of smaller size. This unit also contains boulder-size clasts of black shale and organic-rich dolomitic siltstone. The unit comprises several breccia sub-units, each of which fines upward.

The upper breccia (unit 9) is separated from the Ravenna (black-shale) Bed of the Portwood by a dolomitic, argillaceous sandstone (unit 10), which probably represents a thin Harg Bed. The basal contact of this sandstone is sharp, but it grades up into overlying Ravenna

Figure 11. Folded and overturned mudstone beds (outlined in black dashes) in Unit 7 of the Carpenter Fork Bed just below its contact with the Duffin. White fractured "bed" in the upper part of the photo is an allochthonous clast of Boyle chert at the base of the Duffin.

black shales (unit 11). The Ravenna Bed (unit 11) is similar in lithology to the Carpenter Fork Bed; it consists largely of grayish-black, organic-rich, silty, blocky to fissile shales in fresh exposures and locally contains interbedded dolomitic siltstones.

In the two cores from Powell and Jackson counties, the Carpenter Fork Bed is less than two-meters thick and is overlain conformably by the Harg Bed of the Portwood. In the Powell County core, the contact with the Harg is clearly gradational; in Jackson County, however, this contact is sharp, but apparently conformable. Although dark shales make up the bulk of the bed in each core, greenish-gray shales are present at the base of the bed in the Powell County core. The cores are described in Appendix B.

DISCUSSION

Previous Interpretations

The recognition of some sort of structural influence in the development of the Carpenter Fork section is not new. McFarlan and White (1952) and Rutledge (1957) recognized the presence of faulting on the eastern end of the exposure. Rutledge prepared a geologic map of the area and drew the fault as an arcuate feature trending roughly north-northwest across the eastern end of the cliff face, curving behind the cliff to the north and trending east-west to its westernmost trace about 500-m northwest of the cliff. Displaced beds indicated to him a downthrow of about ten meters on the southwest side of the fault. However, neither McFarlan and White nor Rutledge interpreted the cliff exposure as a graben; rather, they felt that the black shale at the base of the outcrop was a black-shale cave-fill in the Boyle (Figure 12A). In their view, the Boyle was deposited and lithified, a solution cave developed within the Boyle, and the cave was infilled with laminated, black Ohio (Huron-equivalent) Shale. Deformation in the black shale and sagging of the overlying Boyle and Duffin were attributed to load compaction and, possibly in part, to fault movement (Rutledge, 1957).

Lenhart (1985) recognized a fault at the western end of the cliff face in addition to the eastern fault reported by the above workers. Thus, he reinterpreted the Carpenter Fork structure as a graben formed in early Huron time (Figure 12B). Lenhart envisioned an

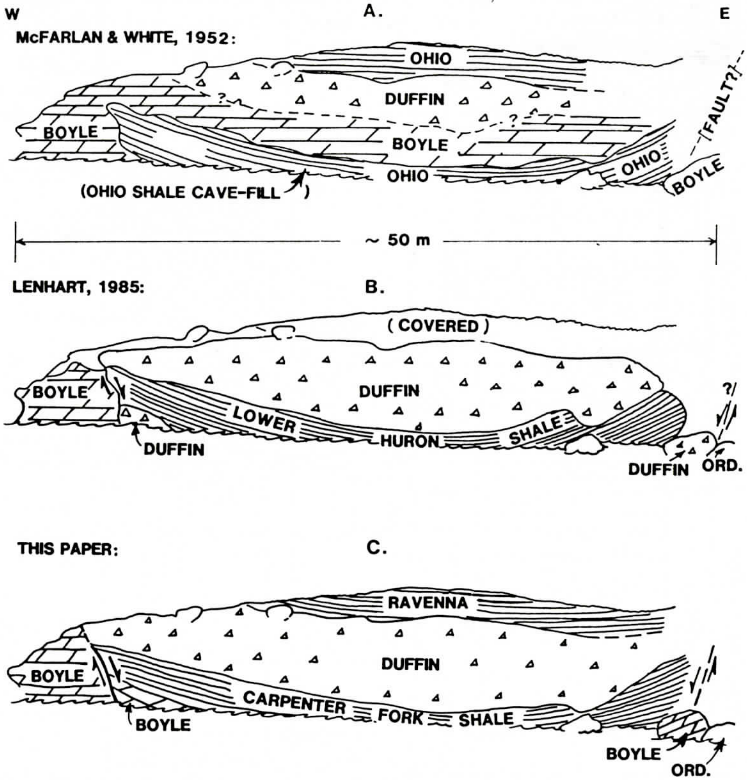

Figure 12. Three interpretations of the Carpenter Fork section: A.) McFarlan and White, 1952; B.) Lenhart, 1985; C.) this study. Horizontally lined areas represent black shales; triangles represent breccias; and the diagonal brick pattern represents dolostones.

allochthonous block or blocks of uplifted Duffin breccia sliding onto the Huron Shale flooring of a downthrown fault block. As evidence for this scenario, he noted the deformed nature of the shale below the Duffin layer, the anomalous presence of two layers of Duffin separated by a few meters of black shale within the graben cut (Figure 12B), and the presence at other locations of penecontemporaneous deformation structures in the lower Huron Shale.

The cave-fill and graben-slide models described above differ from each other primarily because carbonates within the graben were assigned to differing stratigraphic units (Figure 12A, B). As McFarlan and White (1952) interpreted the cliff exposure, the Boyle completely encompasses the Ohio Shale except at the eastern fault boundary, and the Boyle is overlain by Duffin breccias and the Ohio Shale (Figure 12A). Lenhart interpreted as Duffin all the carbonates within the graben, both those above and below the lower black-shale bed (Figure 12A). None of these workers, however, noted that the black shale in the graben was gradational with the underlying Boyle, nor did they closely examine the biostratigraphy, lithology, or radioactivity of the various black shales in the exposure.

Biostratigraphy

Particularly important to interpretations of the graben is the fact that the Carpenter Fork Bed is gradational with the Middle Devonian Boyle Dolostone and is overlain by the Duffin or Harg beds of the Portwood Member, which is also demonstrably Middle Devonian in age (Ettensohn and others, 1989, 1991). Hence, the Carpenter Fork Bed must be Middle Devonian in age and cannot represent parts of the Huron

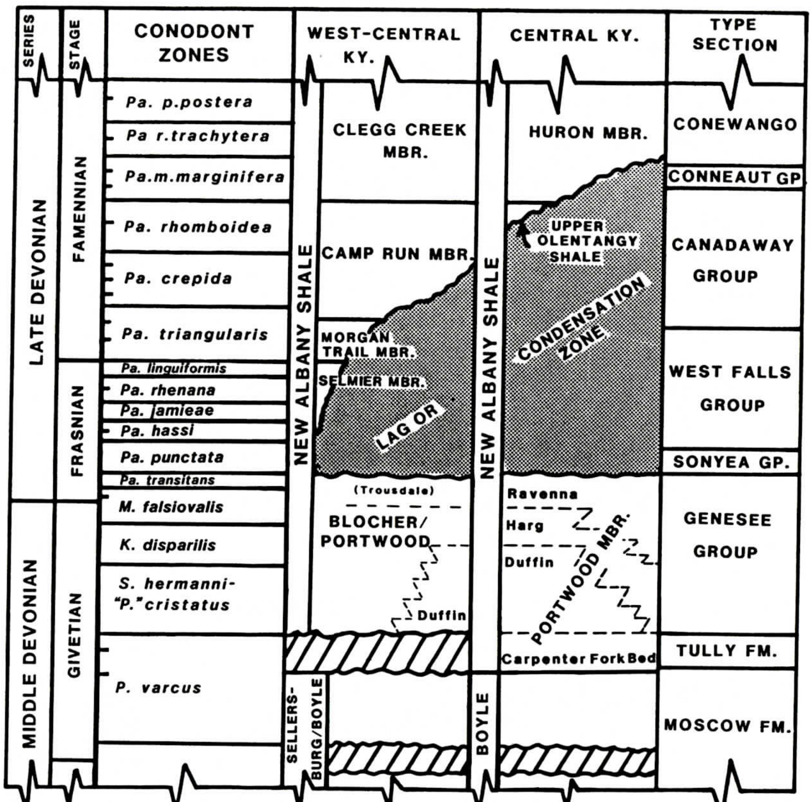

Figure 13. Middle-Late Devonian conodont biostratigraphy and the New York type section compared with New Albany sections in west-central and central Kentucky showing the relative biostratigraphic position of the Carpenter Fork Bed (modified from Ettensohn and others, 1991). Late and late Middle Devonian conodont zones adapted from Ziegler and Sandberg (1984, 1990); remaining Middle Devonian zones from Klapper and Ziegler (1979). *Pa. = Palmatolepis*; *M. = Mesotaxis*; *K. = Klapperina*; *S. = Schmidtognathus*; *P. = Polygnathus.*

Shale Member, which is of Late Devonian (Famennian) age (Ettensohn and others, 1988, 1989; deWitt and others, 1993). In fact, palynology and conodont biostratigraphy suggest that upper parts of the Boyle Dolostone and equivalent units in west-central Kentucky and adjacent parts of Indiana are Givetian (late Middle Devonian) and represent the Lower *Polygnathus varcus* Subzone (Klug, 1983; Wood and Clendening, 1985; Huysken and others, 1992) (Figure 13). In contrast, the various overlying Portwood beds and the equivalent Blocher Member of the New Albany Shale apparently represent the *Schmidtognathus hermanni - "Polygnathus" cristatus* through *Palmatolepis transitans* zones, or latest Middle Devonian through earliest Late Devonian time (Ettensohn and others, 1989) (Figure 13), although our interpretation of the Blocher-Portwood conodont zonation differs from that of Sandberg and others (1994). Some *P. varcus* Zone conodonts are found in the Portwood and Blocher members, but we believe that most of these were probably reworked from underlying Boyle carbonates (Ettensohn and others, 1991). In the intervening Carpenter Fork Bed, the conodont *Polygnathus ansatus*, which is restricted to the Middle *P. varcus* Zone (see Klapper *in* Ziegler, 1977, p. 443) has been found. This particular sample was collected in a sandy carbonate near the base of the Carpenter Fork Bed (Figure 8) and could represent a subtle lag horizon. Many of the conodont elements in the sample are broken with a few showing obvious abrasion. This suggests that at least part of the conodont fauna may be reworked. The free blade on each of eight specimens of *P. ansatus* is broken, which is not too unusual; however, these specimens do not show any obvious signs of abrasion. Therefore, these specimens could be reworked from a younger portion of the underlying Boyle, but we believe that they more likely represent earliest Carpenter Fork deposition. Remaining conodonts (*Polgnathus liguiformis linguiformis, P. varcus*, and *Icriodus cf. I. latericrescens latericrescens*) in this sample are somewhat longer ranging, but are compatible with a late Middle Devonian (Givetian) age. This could mean that parts of the Carpenter Fork Bed are equivalent to the upper Boyle Dolostone elsewhere, or more likely, that the Carpenter Fork Bed represents at least the interval of post-Boyle time corresponding to the Middle *P. varcus* Zone. The Upper *P. varcus* Subzone, which is always difficult to recognize, may also be present in the Carpenter Fork Bed, but it may be unsampled or included in lower parts of the Duffin or Harg beds of the Portwood. Although at present, the conodonts cannot be any more precise in placing all horizons of the Carpenter Fork biostratigraphically, the available conodont data and the fact that all three involved units are gradational or apparently conformable with each other in exposure or cores suggest that the Boyle, Carpenter Fork, and Duffin represent a sequence in natural and reasonably close biostratigraphic continuity as suggested in Figure 13, although minor hiatuses cannot be ruled out with the limited amount of biostratigraphic control.

Other differences between the Carpenter Fork Bed and Huron Shale Member are related to lithology, paleontology, and radioactivity. Dolomitic black shales, dolostones, and shallow-water trace fossils are unknown from the Huron Member. Moreover, radioactivity profiles (Ettensohn and others, 1979) of the Carpenter Fork Bed (Figure 8) show intensities of radioactivity one-third to one-half of that expected from the lower Huron Shale.

Origin of the Carpenter Fork Bed

We agree with Lenhart (1985), for reasons described in a previous section of this paper, that the Carpenter Fork structure is a graben; however, based on our stratigraphic and biostratigraphic analyses, the Carpenter Fork black shales represent a new and wholly autochthonous unit apparently not preserved in many places. We have included the Carpenter Fork Bed in the New Albany Shale because of similar lithology and placed it within the Portwood Member because of stratigraphic proximity, apparently conformable relationships, and similar circumstances of occurrence.

Carpenter Fork shales may have blanketed much of the western Rome trough region or may have been confined to smaller topographic lows, such as sags developed as a result of growth-fault reactivation (Figure 14A). Unfortunately, the complete areal extent of Carpenter Fork deposition is unknown, but a more extensive distribution is probable in light of the common occurrence of black-shale clasts in overlying Duffin breccias throughout the region. The Carpenter Fork Bed is the only likely source for such clasts, and we presume that the absence of the bed in other exposures is due to pre-Duffin erosion.

The origin of the Carpenter Fork Bed is probably related closely to inception of the third tectophase of the Acadian orogeny, because its deposition was nearly contemporaneous with a period of regional subsidence and/or sea-level rise (Tully Limestone and equivalents; Figure 13) accompanying initiation of third-tectophase deformational loading (*e.g.*, Ettensohn, 1985, 1987; Ettensohn and others, 1988). This transgressive event apparently preceded bulge migration and unconformity formation, for no unconformity is present below the Carpenter Fork; yet beyond the Carpenter Fork graben and similar situations, Carpenter Fork black shales are absent and a major regional unconformity (disconformity or angular unconformity) separates the Boyle from the

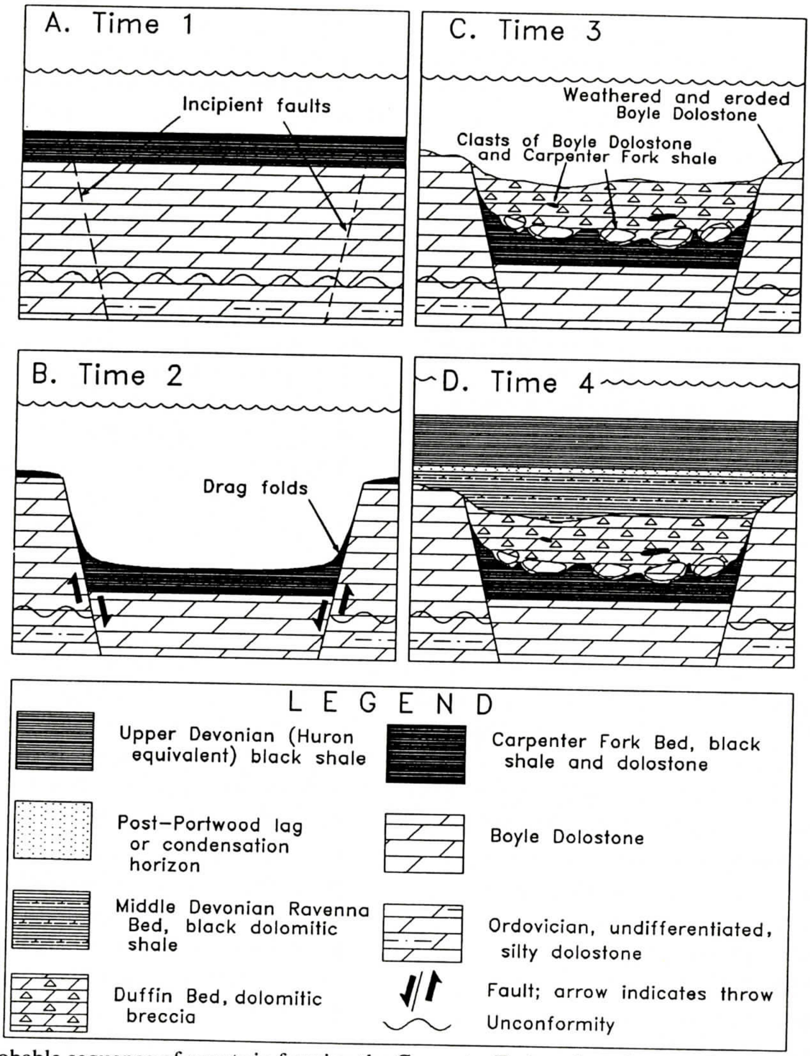

Figure 14. Probable sequence of events in forming the Carpenter Fork graben. See text for description.

Duffin or Harg beds of the Portwood Member (Ettensohn and others, 1991). Assuming a former, more widespread distribution of the Carpenter Fork Bed, this can only mean that regional uplift, structural reactivation, and erosion, probably related to bulge migration through the area (*e.g.*, Beaumont, 1981; Quinlan and Beaumont, 1984; Ettensohn, 1987, 1994), apparently ended Carpenter Fork deposition and resulted in its destruction throughout most of the region. Only in local structural lows like the Carpenter Fork graben, where net subsidence offset the effects of uplift, were the Carpenter Fork black shales preserved. Nearly everywhere else in the region, an unconformity atop the Boyle separates it from overlying Middle or Upper Devonian units (Figure 5A, B) and apparently marks arrival of the bulge along with related uplift and erosion.

The Boyle itself appears to represent a series of largely regressive Middle Devonian (upper Hamilton) carbonates deposited during relaxation phases of the second Acadian tectophase. The unit consists of three to four, relatively thin, disconformity-bound carbonates or sands unconformably overlying Upper Ordovician or Lower Silurian strata in the area (Conkin and others, 1973, 1976; Woodrow and others, 1988). Preserved parts of the Boyle are largely restricted to the Rome trough (Ettensohn and others, 1988), a distribution which may have also characterized the Carpenter Fork Bed at one time. Once transgression began, however, very shallow-water Boyle carbonate deposition gave way to deposition of the somewhat deeper water, organic-rich dolostones and shales of the Carpenter Fork Bed (Figure 14A).

Carpenter Fork black shales were almost certainly *not* deposited in deep, anoxic waters, commonly associated with some of the Upper Devonian New Albany shales (*e.g.*, Ettensohn and Barron, 1981). Although black shales are commonly associated with deeper, anoxic waters (*e.g.*, Demaison and Moore, 1980), great depths and anoxia are not necessary for their deposition (Pedersen and Calvert, 1990). What may have been more important was the presence of abundant organic productivity and the absence of clastic dilution. The Carpenter Fork Bed, moreover, bears evidence of somewhat shallow depositional environments in the form of sedimentary structures (scours, hummocky cross-beds, wavy and lenticular bedforms, and micro-cross-laminae), which indicate that wave base was never too far above bottom, and in the form of abundant trace fossils like *Cruziana*. In fact, association of black shales with bedforms that reflect oxygenating waves and bottom currents and with trace fossils that probably reflect organism consumption of organic matter from bottom sediments, may indicate that production and sedimentation of organic matter simply outstripped the ability of oxygenated water or of organisms to destroy it. In addition, the generally low-lying, carbonate terrain that predominated in the area before Carpenter Fork deposition largely precluded major clastic dilution of the organic matter.

Although we cannot preclude the possibility of synsedimentary fault movement during Carpenter Fork deposition, the absence from the black shales of any coarse clastic debris, which would have likely accompanied fault movement and development of relief in the graben, suggests that major fault movement must have followed or possibly ended deposition of the Carpenter Fork Bed (Figure 14B). Certainly, the drag folds in Carpenter Fork beds along the bounding faults in the graben suggest this. Before the upper parts of the shale had lithified, however, clasts of Boyle from the upthrown fault block either fell or were washed into the graben, deforming the shale below (Figures 10, 11, and 14C). The resulting breccia deposits, consisting largely of angular Boyle clasts, constitute the lower unit of the Duffin Bed of the Portwood (unit 8). Several fining-upward breccia subunits in the upper portion of the cliff exposure (unit 9) indicate subsequent flows of Boyle debris into the graben.

There is little in this exposure to indicate whether the Duffin was a subaerial or subaqueous deposit, although the presence of marine trace fossils in a nearby Duffin exposure supports subaqueous deposition. Much of the debris that was dumped into this or other structural lows or that is found blanketing the Boyle beyond such lows must have come from deeply weathered Boyle substrates, indicating the former presence of substantial relief, exposure and weathering of some parts of the Boyle, all of which were probably related to faulting. The presence of such substrates is borne out locally by paleokarst and weathered residua associated with the Boyle (*e.g.*, Ettensohn and Bayan, 1990; Bayan and Ettensohn, 1994).

The Duffin breccias in the Carpenter Fork graben grade upward through a thin, probable Harg interval (unit 10), into upper Middle Devonian black shales (unit 11) of the Ravenna Bed (Figures 5, 8, and 14D). This sequence marks a progressive deepening with time, and the Ravenna signals the filling of local basins, the end of local sedimentation, and the beginning of regional sediment starvation as is indicated by a thin, sandy, condensation or lag horizon on top of the Ravenna elsewhere in the area (Figures 5C and 14D). This lag horizon is not exposed, however, at the Carpenter Fork section. The condensation horizon contains parts of nine conodont zones (Figure 13) representing nearly ten million years and indicates that, although seas covered the area, little sedimentation, except for winnowing and reworking, intervened (Ettensohn and others, 1989, 1991). Not until the Acadian foreland basin filled in Late Devonian time (Famennian), did muds and silt spill out across the craton and sedimentation resume in central Kentucky with deposition of the Upper Devonian upper Olentangy or lower Huron members (Figures 1 and 13) of the New Albany Shale (Figure 14D).

CONCLUSIONS

Based on stratigraphy and biostratigraphy, the Carpenter Fork Bed is a new — and older — black-shale unit at the base of the New Albany Shale in central and east-central Kentucky. Its gradational contact with the subjacent Boyle Dolostone and apparently conformable position below the Duffin or Harg beds of the Portwood Member of the New Albany indicate a Middle Devonian (mid-Givetian; probably Middle to Upper *P. varcus* Zone or slightly younger) age. Because of lithology and stratigraphic relationships, it is included as the lowermost bed of the Portwood Member of the New Albany Shale.

The Carpenter Fork Bed is known from a single exposure along Carpenter Fork in Boyle County, central Kentucky, and from two core holes in Jackson and Powell counties, east-central Kentucky. Despite its very limited known distribution, apparently restricted to Middle Devonian structural lows or grabens, the unit was probably at one time more widespread, perhaps throughout parts of the western Rome trough, based on the common occurrence of black-shale clasts in the overlying Duffin Bed of the Portwood Member.

Based on lithology and trace fossils, Carpenter Fork black shales, although still representing relatively shallow aerobic to dysaerobic environments, represent an abrupt episode of subsidence or sea-level rise accompanying initiation of the third tectophase of the Acadian orogeny and the end of second-tectophase relaxation reflected in very shallow-water Boyle carbonates. Subsequent uplift of the area along four to five complex fault systems resulted in erosion of the Carpenter Fork Bed and development of an unconformity atop the Boyle Dolostone. The Carpenter Fork black shales are preserved only in structural lows like the Carpenter Fork graben and, hence, only in such situations is there a complete record of Middle to early Late Devonian deposition in central Kentucky and nearby areas.

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APPENDICES

Appendix A: Description of the Type Section of the Carpenter Fork Bed

Creek-cut cliff on north side of Carpenter Fork, east of Carpenter Creek Road (see Figure 7), 1.25-km (4100-ft) south of its junction with Kentucky State Highway 78; Parksville (7½-minute) quadrangle; Boyle County, Kentucky; Carter Coordinates: 2650 ft FSL × 100 ft FWL, 14-M-56.

Name	Description	Thickness (equivalents) Meters	(Feet)
New Albany Shale (incomplete; Devonian):			
Portwood Member (incomplete):			
Ravenna Bed (incomplete):			
	11. Shale, grayish-black (N2), fissile to flaggy; locally contains silt laminae, zones of pyrite nodules, and bodies of brownish-black (5YR 2/1) dolosiltite ----	3.66	(12.0)
Harg Bed:			
	10. Sandstone, dusky-brown (5YR 2/2), dolomitic, argillaceous, laminated; upper unit contains interbedded partings of fissile black shale and very thin beds of organic-deficient, micro-cross-laminated sandstone with dark-shale flaser beds; locally pyritic; contact with underlying Duffin is sharpand marked by quartz nodules; bioturbated ---------------------------	0.15-0.43	(0.5-1.4)
Duffin Bed:			
	9. Breccia; matrix dolarenite, dark yellowish-brown (10YR 4/2), very fine-grained; locally contains sand- to pebble-size phosphatic nodules; porous to vuggy; contains clasts of dolarenite, chert and shale; dolarenite clasts, pale yellowish-orange (10YR 8/6), graded, boulder- to sand-size; chert clasts, light gray (N7), sparse sand- to boulder-size; shale clasts, grayish-black (N2), sparse, up to boulder-size; unit consists of distinct to indistinct subunits that fine upward from coarse breccias to dolomitic mudstone and, locally, shale partings; each subunit thickens toward center of graben --	1.10-4.88	(3.6-16.0)
	8. Breccia; matrix dolarenite, dusky-brown (5YR 2/2) to dusky yellowish-brown (10YR 2/2), silty; vuggy; pyrite nodules present in upper half meter; load structures and flame structures from the underlying shale have deformed the base of the unit; contains clasts of dolarenite, chert, and shale; dolarenite clasts, yellowish-brown (10YR 5/5) to very light gray (N8), composed of porous Boyle dolarenite; graded, boulder- to sand-size; chert clasts, light gray (N7), fractured, granule- to cobble-size; clasts appear to be preferentially oriented; shale clasts, grayish-black (N2), pebble-size, occur locally; unit appears to thicken eastward where clasts may be boulders up to a few meters across; basal contact fairly sharp and highly irregular --	0.43-1.62	(1.4-5.3)
Total thickness of Duffin Bed --		1.53-6.50	(5.0-21.3)
Carpenter Fork Bed:			
	7. Mudstone, dusky-brown (5YR 2/2), dolomitic, silty, argillaceous, sparsely glauconitic; contains thin, dolomitic siltstone laminae; unit is intensely contorted, folded, overturned, and thrusted; contains allochthonous chert clasts up to 0.7 m (2.3 ft) in length; may be injected upward into overlying unit as flame structures and hence occurs only sporadically beneath overlying unit---	0.0-0.43	(0.0-1.4)
	6. Shale, grayish-black (N2), fissile, silty; locally contains dolomitic siltstone stringers and laminae as in unit below; truncates underlying unit toward west; locally contains allochthonous chert nodules-------------------------	0.0-0.27	(0.0-0.9)
	5. Dolarenite and shale, interbedded and interlaminated; dolarenite, light brownish-gray (5YR 5/1), very fine-grained, sandy, organic-rich, glauconitic, argillaceous, locally pyritic; exhibits scours, ripples and micro-cross-laminae; shale, grayish-black (N2), fissile; sand and shale exhibit wavy,		

flaser and lenticular bedding and abundant bioturbation in the form of vertical and horizontal burrows (*Cruziana* and escape burrows present); on east end, unit cut by quartz veins; microfaults throughout; on either end of outcrop, unit thins or is truncated by superjacent unit; at east end, upper parts of unit appear to grade into fissile black shale as in unit below-	0.0-0.73	(0.0-2.4)
4. Shale, grayish-black (N2), silty, fissile; locally contains silt laminae and zones of pyrite nodules; locally contains bodies of brownish-black (5YR 2/1), silty, sparsely glauconitic dolosiltite around which lingulate brachiopods are abundant; upper half meter contains an increased number of dolomitic, pyritic siltstone stringers which may exhibit micro-cross-laminae and wavy beds--	3.57	(11.7)
3. Dolarenite, dark brown (5YR 2/2), fine-grained, sandy, organic-rich, argillaceous, thin-bedded to laminated with hummocky cross-beds, scours and thin shale partings; prominent shale bed exhibits wavy and lenticular bedforms at 0.12 m (0.4 ft) below top; near top, lenses and beds of organic-deficient sandstone are also locally present; local accumulations of dolostone clasts near base; spiriferid brachiopods present locally; basal contact gradational ------------------------------------	0.40-	(1.3)
Total thickness of Carpenter Fork Bed---	5.40	(17.7)
Total measured thickness of New Albany Shale---	10.59-15.99	(34.4- 52.45)
Boyle Dolostone (incomplete; Devonian):		
2. Dolarenite, greenish-gray (5GY 6/1), fine-grained, sandy, argillaceous, in very thin and irregular beds; friable; contains sand- to pebble-size phosphate nodules; locally micro-cross-laminated with organic-rich mud flasers; top of unit characterized by one or two silica veins or concretionary zones; burrowed horizons give rise to local reentrants in the unit----------	0.70	(2.3)
1. Dolarenite, medium gray (N7) to yellowish-brown (10YR 5/2), fine-grained, sandy toward top; silty, argillaceous, locally contains yellowish-brown, sand-size dolostone clasts and granule-size phosphatic debris; locally pyritic; locally contains chert nodules, calcite-filled vugs and quartz-filled fractures near faults, especially on eastern end of graben; contains silicified corals; unit below covered in creek bed----------------------------	0.70	(2.3)
Total measured thickness of Boyle Dolostone--	1.40	(4.6)

Appendix B: Description of Cores

Core D-8

Clay City (7½-minute) quadrangle; Powell County, Kentucky; Carter Coordinates: 2775 ft FSL × 400 ft FEL, 6-P-68.

<u>Name</u> <u>Description</u>	<u>Thickness</u> <u>(equivalents)</u> <u>Meters</u>	<u>(Feet)</u>
New Albany Shale (incomplete; Devonian):		
Huron Shale Member (incomplete):		
10. Shale, dark greenish-gray (5GY 4/1) to greenish-gray (5GY 6/1), silty; contains pyrite laminae; bioturbated---	0.24	0.8
9. Sandstone, greenish-gray (5GY 6/1), quartzose; well-rounded, coarse-grained; contains pyrite and phosphate grains; possible shale clasts; basal contact sharp--	0.03	(0.1)
Total measured thickness of Huron Shale Member---	.0.27	(0.9)
Disconformity		
Portwood Member:		
Ravenna Bed:		
8. Shale, brownish-black (5YR 2/1), silty, slightly dolomitic; contains thin interbedded laminae of dolarenite; basal contact gradational--------------	1.1	(3.6)

Harg Bed:

7. Shale and interbedded dolarenite, brownish-black (5Y 2/1), slightly pyritic; bioturbation, especially prominent beneath dolarenite beds; basal contact gradational, picked at top of pyritic lag(?)-- 1.62 (5.3)

Carpenter Fork Bed:

6. Shale, brownish-gray (5YR 4/1), with thin laminae of dolarenite; dolarenite thickens at top of unit; basal contact gradational--- 0.91 (3.0)

5. Shale, dark greenish-gray (5GY 4/1), interbedded with brownish-black (5YR 2/1) shale; slightly dolomitic; bioturbated; basal contact gradational-------------- 0.67 (2.2)

Total thickness of Carpenter Fork Bed-- 1.58 (5.2)

Total thickness of Portwood Member-- 4.30 (14.1)

Total measured thickness of New Albany Shale--- 4.57 (15.0)

Boyle Dolostone (incomplete; Devonian):

4. Dolosiltite, brownish-gray (5YR 4/1), crinoidal, bioturbated-------------------------- 0.12 (0.4)

3. Dolosiltite, grayish-black (N2), shaly; contains crinoid columnals and phosphate clasts-- 0.12 (0.4)

2. Dolosiltite, brownish-gray (5YR 4/1), interbedded with shale, dark brownish-gray (5YR 3/1); bioturbated-- 0.09 (0.3)

1. Dolosiltite, brownish-gray (5YR 4/1); crinoidal; contains chert nodules; bioturbated- 2.26 (8.5)

Total measured thickness of Boyle Dolostone-- 2.59 (8.5)

Core T-14 (110 ft to 165 ft)

Big Hill (7½-minute) quadrangle; Jackson County, Kentucky; 37° 32' 39" N Lat.; 84° 10' 48" W Long.; (equivalent Carter Coordinates: 2000 ft FNL × 1000 ft FWL, 11-M-64).

Name	Description	Thickness (equivalents) Meters	(Feet)
New Albany Shale (incomplete; Devonian):			
Huron Shale Member (incomplete):			
	13. Shale, grayish-black (N2), silty; contains sparse dolomitic laminae------------------	2.07	(6.8)
	12. Shale, greenish-gray (5GY 6/1) and grayish-black (N2); contains pyrite nodules; contains sparse dolomitic laminae commonly bioturbated below black-shale intervals---		
	11. Shale, grayish-black (N2); contains sparse interbedded dolomitic laminae; base of unit marked by sandy lag horizon--	0.61	(2.0)
Total measured thickness of Huron Shale Member ---		6.12	(20.1)
Disconformity			
Portwood Member:			
Ravenna Bed:			
	10. Shale, grayish-black (N2); contains sparse dolomitic laminae and burrow-fills; basal contact gradational---	1.83	(6.0)
	9. Shale, grayish-black (N2) with interbedded dolarenite laminae; basal contact gradational---	0.61	(2.0)
Total thickness of Ravenna Bed---		2.44	(8.0)
Harg Bed:			
	8. Dolosiltite, greenish-gray (5GY 6/1), bioturbated; basal contact gradational---------	0.45	(1.5)
	7. Dolarenite, light olive-gray (5Y 6/1), fine-grained; thin-bedded at base, laminated in upper unit; contains interbeds of grayish-black (N2) dolosiltite and shale near top; dolosiltite beds in upper 0.15 m (0.5 ft) intensely bioturbated; basal contact sharp, irregular--	1.49	(4.9)
Total thickness of Harg Bed--		1.94	(6.4)
Carpenter Fork Bed:			
	6. Shale, grayish-black (N2), poorly fissile, very slightly dolomitic; contains laminae to very thin beds of very fine-grained dolarenite; sparsely burrowed, but upper 0.12 m (0.4 ft) is intensely bioturbated; basal contact gradational-------	1.07	(3.5)
Total thickness of Portwood Member---		5.45	(17.9)

Total measured thickness of New Albany Shale	11.57	(38.0)
Boyle Dolostone (Devonian):		
5. Dolosiltite, light olive-gray (5Y 6/1), silty; lenticular and flaser beds; locally contains crinoid ossicles and dolomite-filled vugs; highly bioturbated; contains minor greenish-gray (5GY 6/1) and brownish-black (5YR 2/1) shale interbeds; upper 0.03 m (0.1 ft) becomes increasingly silty; basal contact sharp, irregular, possibly bioturbated or scoured	1.01	(3.3)
4. Dolosiltite, light olive-gray (5Y 6/1), crinoidal; bedding obscure; horsetail stylolites common; extensively bioturbated; upper 0.5 m (1.5 ft) of unit marked by silty-shale zone; basal contact sharp, irregular, possibly bioturbated or scoured	1.19	(3.9)
3. Dolosiltite, similar to unit 5; basal contact sharp, stylolitic	1.07	(3.5)
2. Sandstone, medium gray (N5), well-rounded coarse quartz grains; dolomitic; contains pyrite nodules, granule-size clasts of sandy mudstone and greenish shale; basal contact sharp, irregular	0.06	(0.2)
Total thickness of Boyle Dolostone	3.33	(10.9)
Unconformity		
Crab Orchard Formation (incomplete; Silurian):		
1. Shale, medium greenish-gray (5G 5/1), silty, crumbly, plastic when wet; locally dolomitic; locally contains glauconite and pyrite bands; upper 0.03 m (0.1 ft) bioturbated, filled with sands from unit above; end of core	1.92	(6.3)

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GEOMORPHIC DEVELOPMENT AND PALEOENVIRONMENTS OF LATE PLEISTOCENE SAND HILLS, SOUTHEASTERN LOUISIANA: DISSCUSSION AND REPLY

DISCUSSION

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A paper by Mossa and Miller (1995) recently proposed the erosional origin of scores of small hills on the 50 km (c. 31 mi) wide Prairie surface in the Florida Parishes of southeast Louisiana. Due to his untimely passing more than eight years ago, Bob Miller, an eminent, sadly missed Louisiana pedologist of great promise did not participate in the evaluation of the more recent literature and preparation of the paper. For this reason, I am referring only to J. Mossa in the following.

Several early publications that deal with well preserved, elongated ridges on the Late Pleistocene alluvial coastwise Prairie surface agree on their eolian origins. The dune trends represent the southernmost sector of the discontinuous Mississippi Valley Pleistocene dune belt, recently mapped by Saucier and Snead (Plate 6; in: Autin and others, 1991).

In denying eolian ridge origins, Mossa mentions sediment characteristics, present soil types, and drill data. Not specified, her drill samples probably came from short push-bored cores or augering. This discussion is hampered by the fact that Mossa does not provide specific details about any proposed erosional mechanism and the overall regional stratigraphic-geomorphic relationships in arguing her ridge-forming concept.

THE ISSUES AT HAND

The evidence for eolian dune or fluvial-erosional origins in part falls into three categories.

{1) Dune ridge morphology

Scores of small hills, here considered relict dunes, are elongated oval-elliptical bodies, oriented mostly from NNW-SSE to N-S directions. The largest ones reach 8-to-10 m above the surrounding plain and are bounded by high, steep slopes. The hillocks are elliptical in map view (Otvos, 1971, 1991, and 1995). An additional large parabola dune occurs near Denham Springs, La. (Otvos, 1971). I am unaware of any geological process capable of carving these symmetrical, sharply defined landforms by fluvial erosion; exclusively out of well-sorted medium sand bodies of fluvial origin, at that.

Other landforms; irregularly shaped, flat-topped interfluves of low relief in the Skulls Creek area (E 1/2 of sec.5, R8E, T5S; Mossa's Fig. 3), did indeed form by shallow fluvial incision of the alluvial Prairie surface. A flat-topped, low feature, west of a real dune ridge (Fig. 5) was one of Mossa's only two documented "sand hill" examples. Unoriented, irregularly shaped, relatively low interfluve mounds that barely rise above the surrounding plain were carved by surface dissection from sandy (-pebbly) Prairie alluvium. They appear to be the "many hills" that, according to Mossa are gravel "mines".

Mossa's objection to the dune origin of the hills is also based on the fact that pebble-rich sand layers that underlie the elongated ridges occur at elevations higher than the surrounding alluvial plain. However, this configuration is due to the ridge position on the interfluve. The pebbly alluvium, covered by dune ridges may occur higher than the elevation at which it forms the adjacent lower slope of the interfluve: where not sheltered by dune ridge, slope erosion did lower the alluvial deposits.

(2) Sedimentary textures and structures

Good sorting is typical of these dune ridges. Excluding illuvial and paleosol mud components that occur as dark mottled fossil soil lenses, pods and nests (Fig. 2 in Otvos, 1971), the sand granulometry, including sorting, is compatible with dune origins. Silty-clayey layers within the eolian sand lithosomes are of pedogenic, not alluvial origin. These structures and the elimination of eolian lamination by abundant root and other biogenic activity are characteristic of certain eolian deposits.

Similarities between the granulometric parameters of a thin, well sorted, steeply cross-stratified (megarippled) fluvial sand interval at Slidell, La., cited by Mossa after Otvos (1971), and of the dune sands do not prove fluvial origins of ridge lithosomes. Sand ridges are absent from the Prairie surface in the entire region that surrounds this borrow pit (Otvos, 1971, p. 1755; Mossa and Miller, p. 81) and the fluvial nature of its deposits was never in question.

Had Mossa distinguished between the sand and gravel fractions and illustrated variations of granulometric parameters in the sections (Fig. 6), the vertical separation between eolian dune and floodplain deposits would become obvious. The several m thick, pebble-free sand deposits that directly underlie the typical dune ridges, if well sorted, may be of point bar or channel bar origin.

3) Unconformity between ridge-forming and underlying deposits?

Mossa's drill samples may have been insufficiently well preserved for recognition of a clearcut unconformity at the dune base. She does not indicate any unconformity between the depositional units, even if minor breaks are common in alluvial sequences that often are separated by erosional events. These would be expected between pebbly channel and sandy overbank/point bar units. No well-defined unconformity would be expected between the alluvial and eolian lithosomes if the dunes accumulated gradually on the foundation of well sorted alluvial sands with a granulometry that greatly resembled that of dune sands. Root- and other biogenic processes that obliterated the primary sedimentary structures may have also weakened the contrast that existed between the alluvial and dune units.

LOESS AND THE SAND SOURCES

Mossa's reasons for citing dune ridge formation in the context of present Florida Parishes streams and the Late Pleistocene loess cover development remain obscure. The narrow local floodplains and the dominantly sandy sediment load clearly played no role in loess distribution during drier Late Wisconsin intervals. Loess, derived exclusively from the Mississippi River floodplain to the west, would have not interfered with dune genesis either. It is obvious that the present streams "are not a major sediment source for dune creation". Dunes did not form under the humid-subtropical conditions of the Holocene. The shifting-overlapping Sangamonian Interglacial Prairie floodplains did not resemble the present ones, still confined to incised valleys and separated by broad, high interfluves.

Under the right climatic conditions, dune formation would have favored broad, coalescing floodplains in unincised valleys that were wider than the present ones. These floodplains may have been the sand source during (seasonally?) drier climate episodes, not yet documented in the local geologic record. Their formation would have well preceded the full-glacial late Wisconsinan sea level drop and severe stream incision phase, accompanied by generally cooler, windier, and drier climate conditions than those that existed in the Florida Parishes during the Sangamonian and the Holocene.

IMPLICATIONS OF THE "FLUVIAL-EROSION" HYPOTHESIS

Mossa does not address wide-ranging regional implications of her assumption. First of all, fluvial incision presupposes a minimum 10 m (33 ft) vertical aggradation of the Prairie surface (Fig 2). Further sea level rise to unrealistically high Sangamonian levels would have been required during continued alluvial aggra-

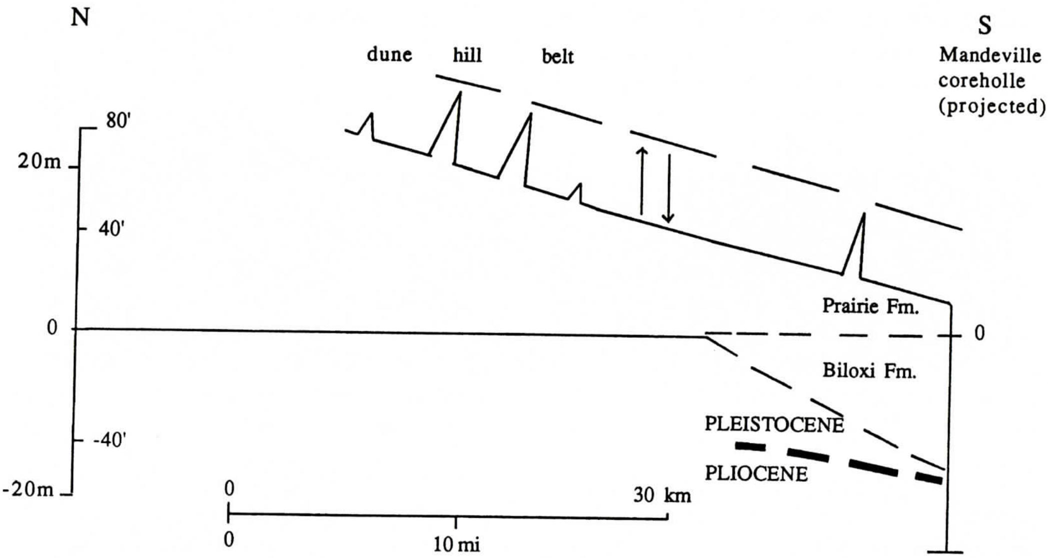

Figure 1. Schematic N-S cross section across central Florida Parishes, La. Vertical arrows point to minimum level of Late Pleistocene aggradation (projected land surface: dashed line) and subsequent surface degradation to accommodate Mossa's premise of fluvial-erosional ridge formation. Mandeville corehole is located SE of the dune area (Otvos and Howat, 1992).

dation to result in associated rise of the regional base level. The Prairie surface would have been raised above the present sand hill summits, thus doubling or even tripling the known thickness of the Prairie Formation (Fig. 1), established in our Mississippi Coast and Fontainebleau State Park (Mandeville, La.) rotary coreholes in adjacent areas.

Fontainebleau Park southeast of Hammond (Mossa Fig. 3) is located on the southern margin of the Prairie coastwise terrace (Otvos and Howat, 1992). Very stiff, dark greenish gray-grayish green, fossil-free sandy muds and sandy clays, correlative of the Pliocene inshore-paralic Graham Ferry Member of the Pensacola Formation (Otvos, 1994) underlie the Pleistocene in the State Park corehole (Fig. 1). In the entire coastal region the Prairie Formation overlies the Sangamonian Biloxi Formation that was deposited in marine-to-inshore facies (Fig. 1; Otvos,1991).

A subsequent, extensive sea level drop would be required to incise the Prairie surface and mold the sand hills in the process. There is no evidence for a commensurate, rapid early Wisconsinan sea level decline and stream downcutting. The incision would have resulted not only in uneven degradation but also intensive dissection of the Prairie land surface. Were the premise of fluvial-erosional ridge sculpting a realistic one, the process would have left hundreds of residual sand ridges scattered across the entire north-central Gulf coastwise Prairie surface. Fluvial incision processes were much more intensive during the subsequent full-glacial record sea level drop and accompanying, still ongoing slow rise of land. Uplift raised the landward fringe of the Prairie surface in Louisiana and adjacent Mississippi at least as high as +24 m (c. +80 ft).

Even despite its most severe, full-glacial incision, the Prairie coastwise surface remains remarkably level and relatively undissected. Outside the discussed and relatively limited dune hills area, it is lacking in prominent hills of this kind. Groups of high Late Pleistocene dune hills occur only at a few other locations; all of them on the present Gulf shore between the south Texas and Florida's Apalachicola coasts (Otvos, 1991, 1992 and 1995). These Wiscon-

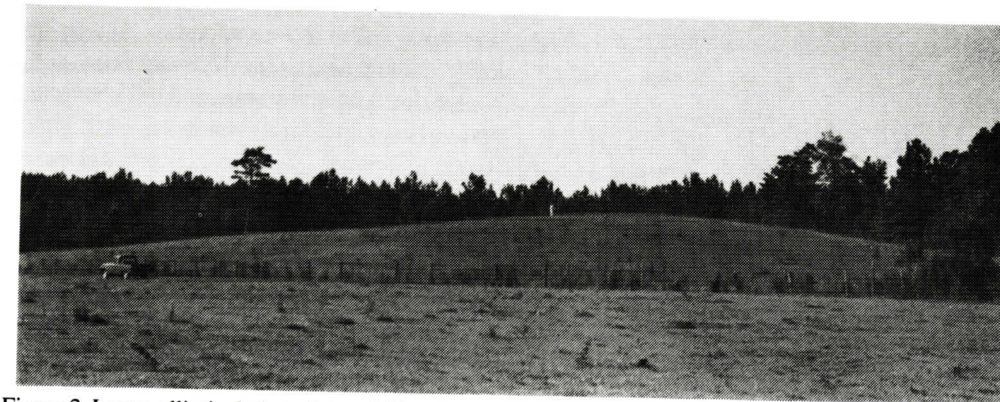

Figure 2. Large, elliptical-shaped dune hill on Prairie Formation alluvium of SE Louisiana (Hoggatt Hill, sec. 21 T5S, R8E, Tangiphoa Parrish).

sinan dunes developed well inland of the corresponding low sea level shoreline. Under relatively dry conditions did intensive deflation of the broad, thinly vegetated Sangamonian Gulfport barrier ridges take place. These barriers represented the dune sand sources (Otvos, 1991, 1992).

One more argument against the "erosional" origins of the southeast Louisiana dune ridges is the presence of relict meander loops imprinted in the southwest Mississippi Prairie surface (Otvos, 1995). Had the Prairie surface undergone erosional degradation, the meanders would not have been preserved.

CONCLUSIONS

Drill data from the southern margin of the Louisiana Prairie coastwise surface and its eastward continuation in Mississippi revealed the presence of marine and inshore/nearshore deposits, members of the Sangamonian Interglacial marine cycle (Otvos and Howat, 1992 and Fig. 1). The Biloxi thus defines the thickness of the overlying and interlayered, regressive alluvial Prairie Formation on the north Gulf coast.

Extensive Late Pleistocene aggradation of a thick Prairie fluvial sequence and subsequent rapid fluvial surface degradation of the Prairie surface would be only one of several prerequisites for the assumed ridge-sculpting by fluvial erosion. Field data, local stratigraphy and Late Pleistocene sea level history do not conform with an inferred intensive fluvial incision of the Prairie surface either during early- or mid-Wisconsinan times.

In addition to sediment and geomorphic data, including also hill orientation and ridge group alignment, stratigraphic constraints provided additional solid support for the eolian genesis of the sand ridges. The crisply outlined, oval ridge shapes could not have been molded by any known process of fluvial erosion during downcutting, or otherwise (Fig. 2). The ever-fluctuating global sea level remained substantially lower than at present throughout the time between the Sangamonian peak and the Late Wisconsinan full-glacial low (e.g., Toscano and York, 1992). Specific sea level positions and ages of potential dune-generating dry episodes in this time interval are known only poorly or not at all.

In discussing local soil categories at great length, Mossa apparently recognizes the futility of using them in mapping Pleistocene stream channel belts. Her generalized mapping of gravel pit areas and associated "sand hill complexes" in Tangipahoa Parish (Fig. 3 of Mossa and Miller) did not convincingly identify specific buried stream channels or channel belts.

Mossa's "broad- brush" approach also led to the inclusion of a sizable area of the much older. Pliocene, Citronelle Formation into the Late Pleistocene "Prairie Terraces" sand ridges and abandoned channels- map categories W and

NW of the town of Amite (Fig. 3). This heavily dissected tract rises above +60 m (c. +200 ft) elevation.

With the exception of the cross-cutting stream valleys, Prairie alluvial deposition in the Florida Parishes ended in early Wisconsinan times. Unlike present local streams, the Sangamonian Prairie floodplains occupied unincised, broad, and shallow valleys along which the well-defined dune hill trends probably developed. The conspicuous zonal alignment of the Florida Parishes sand hills indicates their likely locations along the margins of floodplain- filled valleys.

The absence of reliably documented and dated fluvial deposits outside the present local floodplain areas indicates that the lower-than-present sea levels throughout the Wisconsinan, reinforced by slow regional uplift, had precluded alluviation on the Prairie coastwise surface. The failure to prove (and date) Middle-to-Late Wisconsinan Prairie alluviation, refuted the existence of a hypothetical Late Wisconsinan "Farmdalian upper chronostratigraphic unit"; the suggested upper division of an advocated "Sangamonian-Farmdalian Prairie Complex" in coastal Louisiana (Autin and others, 1991; Otvos, 1991).

Limited erosional incision and dissection of the Prairie surface areas continued at varying rates throughout the Wisconsinan. A record sea level drop rapidly increased coastal plain stream gradients. Following the very brief Farmdalian Interstade (c. 28-to-25 ka B.P.), characterized by relatively higher sea level that remained still well below the present (Otvos, 1991; York and Toscano, 1992). Stream downcutting culminated in late, full-glacial Wisconsinan times, c. 22-to-18 ka B.P.).

An alternate Late Wisconsinan scenario of dune generation {Otvos and Howat, 1992) involves significant episodic reduction in precipitation and related intense seasonal deflation of poorly vegetated, temporarily barren surfaces and of the narrow sandy floodplains.

There is a pressing need for effective new sediment-dating methods in resolving numerous problems that plague Pleistocene coastal stratigraphy. Credible age-dating of several key northern Gulf coastal surfaces and underlying alluvial and brackish sediment units would be a revolutionary step forward.

REPLY

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I thank Otvos (1995) for his interesting comments regarding our paper (Mossa and Miller, 1995) on sand hills and landscape development in southeastern Louisiana. His observations and thoughts are always provocative, even though we are not always in agreement. I was especially pleased that he acknowledged the accomplishments of my deceased eminent co-author Bob Miller, who could not participate in the final phases of this research. Yet, I wish to make clear to Otvos and all others that our opinions were shared ones and had been discussed extensively. Prior to his death, we mutually presented our initial data and interpretations at a professional field trip and at a national conference (Mossa and Miller, 1986a; 1986b). In addition, this field guide was later peer-reviewed and printed for wider distribution (Mossa and Autin, 1989). The interpretations regarding landscape development since these works are essentially unchanged, and only the writing, analysis and presentation have become more sophisticated. Having the utmost respect for Bob Miller as a person and as a scientist, I wish to give proper credit for his efforts.

In this reply, I hope to clarify the areas of agreement and disagreement with Otvos regarding the sand hills in southeastern Louisiana. I agree with Otvos that Bob Miller is sadly missed, that there are many questions left unresolved concerning landscape development of the Gulf Coastal Plain, that a number of the features are fluvial in origin, that at times we did not provide specific details about the timing and nature of erosion, and about the pressing need for more effective dating. However, there are also a number of areas where we largely disagree. I will address most but not all of his comments, as many questions he raised were discussed in our paper and could be answered with a more careful reading of the manuscript. I encourage all interested parties to review our papers, both sets of comments, and all other available information in order to form their own opinions about the origin and development of the sand hills in southeastern Louisiana.

My reply to his comments has several components. First, I will discuss criticisms of a technical nature which include incorrect extrapolation from other papers, careless reading of our paper, inadequate use of secondary resources such as topographic, geologic maps, and soil surveys, and fallacious assumptions regarding our techniques. In addition, I will discuss one comment with technical merit. Second, I will summarize agreements and differences of a topical nature, concerning professional opinions about landscape development. Third, I will discuss areas where we have differences in professional style, specifically the varied approaches that we have taken toward speculation. Finally, I will build further on some comments made regarding the need for research on inliers and hillslope evolution, research areas that may lead toward eventual resolution of some of our differences.

TECHNICAL ISSUES

First of all, these sand hills do not, as Otvos states, "represent the southernmost sector of the discontinuous Mississippi Valley Pleistocene dune belt, recently mapped by Saucier and Snead...in Autin and others (1991)". The southernmost extent of the dunes mapped by Saucier and Snead (1991) falls in central Arkansas, about 440 km or 4 degrees latitude to the north of the study area in southeastern Louisiana. The dunes in Arkansas were formed from valley train deposits of the Mississippi, with a much larger source area and in a very different setting from our study area. A reader not familiar with the map of Saucier and Snead (1991) and writings of Autin and others (1991) might inadvertently conclude that these researchers depict dunes in southeastern Louisiana, and that our study disagrees with their general synthesis. Neither of these is correct. Otvos extrapolated from their works in a manner that they would likely dispute.

Should Otvos have read our paper more carefully, he would have found clear answers to a number of his imprudent, sometimes derogatory, comments. For instance, he states "*Not specified*, her drill samples probably came from *short* push-cored bores or augering". In fact, we clearly stated "Crests, sideslopes, and toeslopes of several hills and lower relief ridges were bored with a *truck-mounted Giddings hydraulic probe*" (Mossa and Miller, 1995, p. 87). Specific information regarding core lengths are described in the same paragraph, and some are depicted in diagrams. Further, the coring techniques we used are widely respected and have been used by soil scientists, geographers, geologists, and archeologists in a number of published studies (e.g. Autin, 1992; Collins et al., 1994; Farrell, 1987; Holliday et al., 1994; Knox, 1987; Mandel, 1992; Mason et al., 1994; Van Nest and Bettis, 1990).

In addition to such errors, a number of negative comments could have been avoided with more attention to topographic and geologic maps and soil surveys. For instance, should he have examined topographic maps more carefully he would have known that the elevations of the areas we mapped did not include "a heavily dissected tract" that "rises above 60 m (200 ft) elevation". In fact, the areas where late Pleistocene soils are depicted on the Albany and Hammond quadrangles are less than 38 m (125 ft). The highest elevations occur at the northern end of the study area, and correspond with regional gradients of the Prairie Complex. Should Otvos have assessed geologic maps more carefully (Snead and McCulloh, 1984), he would have seen the area is mapped as late Pleistocene Prairie Terraces and did not include "a sizable area of the much older Pliocene Citronelle Formation". The map included in our paper was based on field mapping by soil scientists (McDaniel and others, 1990), which show that the soil series mapped on the Citronelle Formation differ from those mapped on late Pleistocene surfaces. In addition, his comments regarding the hills that "*according to Mossa* are gravel mines were not, as he implies, construed in my imagination, but were derived from topographic maps and soil surveys.

Otvos similarly chooses to use disparaging comments when our findings discredit the eolian dune origin of the sand hills. He states "drill samples may have been *insufficiently well preserved* for recognition of a clearcut unconformity at the dune base." In fact, the Giddings probe allows one to examine borings in the field, so preservation is not an issue in recognition of stratigraphic boundaries and soil horizons. A major textural unconformity from a sandy channel deposit to a overbank clay would have been easy to distinguish. It was not, however, encountered. We did bring samples back to the laboratory and, unlike Otvos assumes, took care in wrapping and handling of samples.

One technical comment that Otvos proposed deserves some mention and consideration, namely the granulometric separation of the sand and gravel fractions. Given that the distribution of gravel was sporadic, we generally did not separate these because we thought its distribution was too irregular to be representative. In hindsight, we could have included this data, and cautioned our readers to be wary about the potential variability of this size fraction in particular. I do not concur with Otvos, who claims that if pebble-free sands overlie deposits with gravel they are eolian and that if they underlie deposits with gravel they could be either eolian or fluvial. Although large gravel rules out an eolian origin, stratigraphic position does not distinguish an eolian from a fluvial deposit.

TOPICAL ISSUES

Many of the topical differences between our landscape development model and Otvos' model are best examined by a review of the original papers. There are, however, new areas where we concur and some issues yet unresolved. It appears that we are now in agreement that many of the features in the Skulls Creek area were deposited by ancestral rivers and were modified by incision (Mossa and Miller, 1995; Otvos, 1995), amongst other factors. I consider this progress and credit Otvos for becoming more specific about which hills he considers dominantly fluvial or dominantly eolian. Our paper (Mossa and Miller, 1995) also suggests that this area is not uniform, with differences in the degree and nature of erosional modification and of sediment infilling from the northern to southern end of the sand hill complexes. Given the large number of sand hills and sand ridges in the area, Otvos and I both agree that it is difficult to sample them all and that it is possible that there is some variability in the depositional environment and geomorphic history of different hills across the region.

I also recognize that some of our differences in opinion may have formed from having sampled different sites. He acknowledges that the feature that we sampled and characterized may be fluvial in origin, by commenting that it is "west of a *real* dune ridge", depicted in our Figure 5 (Mossa and Miller, 1995, p. 87). Unfortunately, we were under the impression that he would have considered both features eolian, since they have similar sandy soils and maximum elevations or landscape positions, even though they differ morphologically. Because most of these sand hills occur on private property, dominantly homesites, access depends on permission from landowners. At the time we were sampling, the landowner or tenant of the site that Otvos labels "a real dune ridge" was uncooperative, and it could not be included in our study.

We also are in agreement that local floodplains did not influence loess distribution, that loess from the Mississippi River would not have interfered with dune genesis and that present streams are not a major sediment source for dune creation. Otvos and I concur, in addition, that if any of sand hills are eolian in origin that dune formation would have favored broad floodplains and climates must have been much drier.

We also have areas of disagreement. For example, Otvos states that our results double or triple the known thickness of the Prairie Complex, which he characterizes from borings along the Mississippi coast and Fontainbleau State Park. However, geologic units are not uniform in thickness and have much spatial variability, which depends on initial landscape position, the scale of the source, and the depositional environment, amongst other factors. In fact, local thicknesses of the Prairie Complex, as far as we have determined, fall well below that of sections described in the literature. For instance, Autin and others (1988) described a section of late Pleistocene Prairie Complex channel fill deposits of the Mississippi River which exceeds 20 m in thickness.

Although Otvos has derived opinions about sand hill genesis from observations of the "Biloxi Formation and correlatives to the Pensacola Formation in southeastern Louisiana", neither of these units is currently recognized or mapped by the Louisiana Geological Survey (Snead and McCulloh, 1984). I have not been to these field sites to form my own opinions about the relationships of these to sea level fluctuations and am hesitant to discuss the potential regional implications of a limited number of borings with which I have no first-hand assessment.

Regardless of our differing experiences, however, I do not believe, like Otvos presupposes, that an extensive sea level drop is required to form erosional inliers. Incision is only one of many potential causes of remnant formation. Some other mechanisms for the formation of erosional remnants include avulsions, floodplain modification during overbank events, and stream headcutting, processes that can occur at high sea level stands as at present. Also, I not do presuppose, as he does, that unrealistically high sea levels would have been required for alluvial aggradation. Aggradation depends on relative, not absolute, elevation, and does not only occur beneath marine waters. Spatial variations and timing of cut and fill in the region are complex phenomena, and much more work remains to evaluate these phenomena as well as their relationships to sea level changes.

SPECULATION: AN INDIVIDUAL CHOICE

As Otvos claims, specific details about the proposed erosional mechanisms, timing, or distribution of these features were not provided. This is because the specific details regarding the evolution of these features are unresolved, and we chose not to speculate extensively where we did not have definitive answers (We is used here because this approach was largely inspired by Miller). We believe that our model can be further refined through additional geomorphic and stratigraphic studies and dating, and wish to promote future studies. Our approach differs from Otvos' who prefers to speculate more about conditions and timing of events in the Pleistocene. For instance, he states that "a yet unidentified dry climate episode during late Sangamonian (earliest Wisconsinan?) alluviation associated with the marine highstand, is believed to have been the most likely time of dune formation along broad, coalescing floodplains" (Otvos, 1995).

We generally did not speculate far beyond available evidence and, instead, encourage future researchers to consider multiple working hypotheses. For example, we bracketed the deposition of ancestral channels more broadly than Otvos through use of the loesses and proposed several ideas about the possible nature of ancestral fluvial channels. Further, it is possible that there are additional possibilities that we did not consider. I believe both our and Otvos' approaches concerning speculation has its inherent advantages and disadvantages, and the approach taken should be left to the discretion of each researcher.

FUTURE RESOLUTION?

Recent interactions through this forum have clarified some areas of agreement and disagreement since publication of the Otvos' (1971) first paper. More efforts are necessary toward resolving some of our differences. Otvos suggests we need new sediment-dating methods, yet I believe that more work can be conducted before new techniques are developed. One area of needed research concerns inlier geomorphology and landscape development (Mossa and Miller, 1995). For instance, when discussing the oval, elliptical, and parabolic shapes of some of sand hills, Otvos states "I am unaware of any geological process capable of carving these sharply defined landforms by fluvial erosion". A comprehensive examination of remnant geomorphology in the coastal plain is needed to assess the range of shapes of features carved by fluvial erosion.

Given that Otvos and I are in disagreement as to whether and which of the high-relief sand hills are inliers or dunes, and that these and their surrounding landscapes are currently mapped within the same geologic unit (late Pleistocene Prairie Complex), it would be most useful to begin by examining the shapes and sizes of features that are widely recognized as inliers (i.e. mapped as different geologic units). If the inliers are large enough, such features may be depicted on geological maps. Smaller inliers are generally not depicted because most efforts at geologic mapping have been at a scale of 1:250,000 and smaller.

A number of erosional remnants occur in the southeastern United States, not far from the study area. Some of the largest remnants, which include Macon Ridge and Crowley's Ridge, are in the Mississippi River alluvial valley (see Autin et al., 1991, p. 548 or Saucier, 1994, p. 8). Other remnants described in the literature are located in the Red River valley (Snead, 1993). The Geologic Map of Louisiana depicts these and others, including an inlier in Amite River Valley near Grangeville mapped as the High Terraces (Snead and McCulloh, 1984) or Upland Complex. Several late Pleistocene erosional remnants occur in the marginal basin near Lakes Pontchartrain and Maurepas, and along the Pearl River valley and mouth (Snead and McCulloh, 1984; Saucier, 1994). Even at a small scale, some are mappable (Fig. 1, generalized from 1:500,000 maps). Given the scales used for mapping, and the limited number of remnants associated with other geologic units, Otvos' expectation of "hundreds of residual sand ridges across the Prairie surface", is unreasonable.

Mappable remnants have a range of sizes and shapes, and their form is in part depositional and in part erosional. Following deposition, size and form can be modified by a variety of processes which include incision, headcutting, avulsion, secondary channel development, tributary dissection, marine processes, and various combinations of these. New remnants may be forming where the zone of contact with the contiguous geologic unit is small (Fig. 1). Headward erosion appears to be important in their development.

The nature of landform development would vary with geology and vegetation. Erosional forms carved in dominantly sandy sediments would differ appreciably from those in loamy, silty, or clayey sediments. Initial studies could begin by comparing the dominantly silty-clayey remnants of the Prairie Complex with the dominantly sandy remnants of the Deweyville Complex (Fig. 1). The degree of sediment consolidation and cementation and the nature of vegetation cover would affect erosional patterns and the ultimate form of the remnant. Resolving the sand hill controversy will require study of a number of remnants with similar sediment characteristics to these features.

Also, more research concerning hillslope evolution in our study area is necessary, and some general points for discussion can be illustrated using 1:250,000 Digital Elevation Models of the study area. The contours on the maps from which the models are derived and the subsequent digital representations are highly generalized and do not show the sand hills and a number of other similar size topographic features. Further, the 1:250,000 maps contain

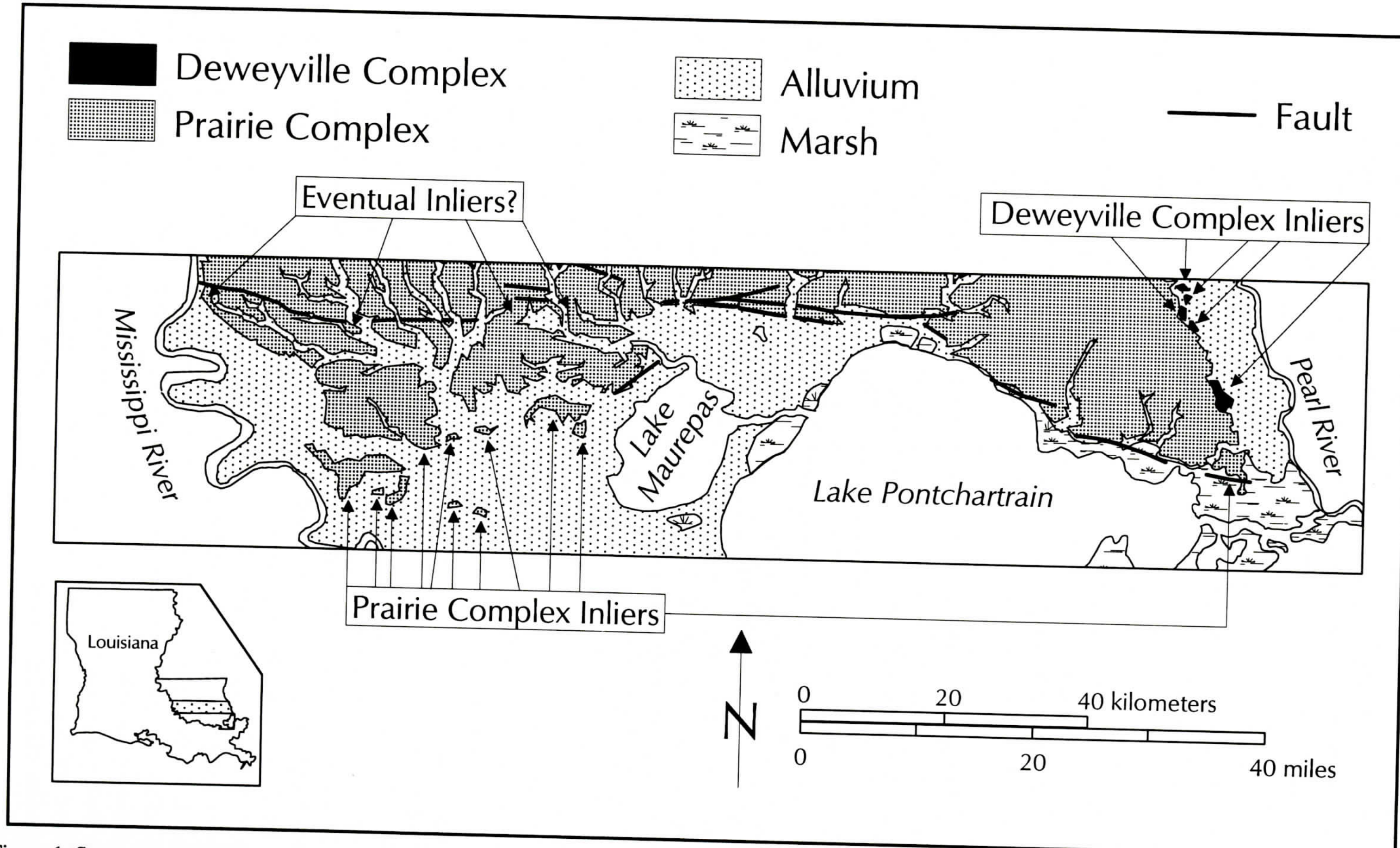

Figure 1. Some late Pleistocene erosional remnants in southeastern Louisiana are mappable at a small scale. Geology generalized from the Geologic Map of Louisiana (Snead and McCulloh, 1984).

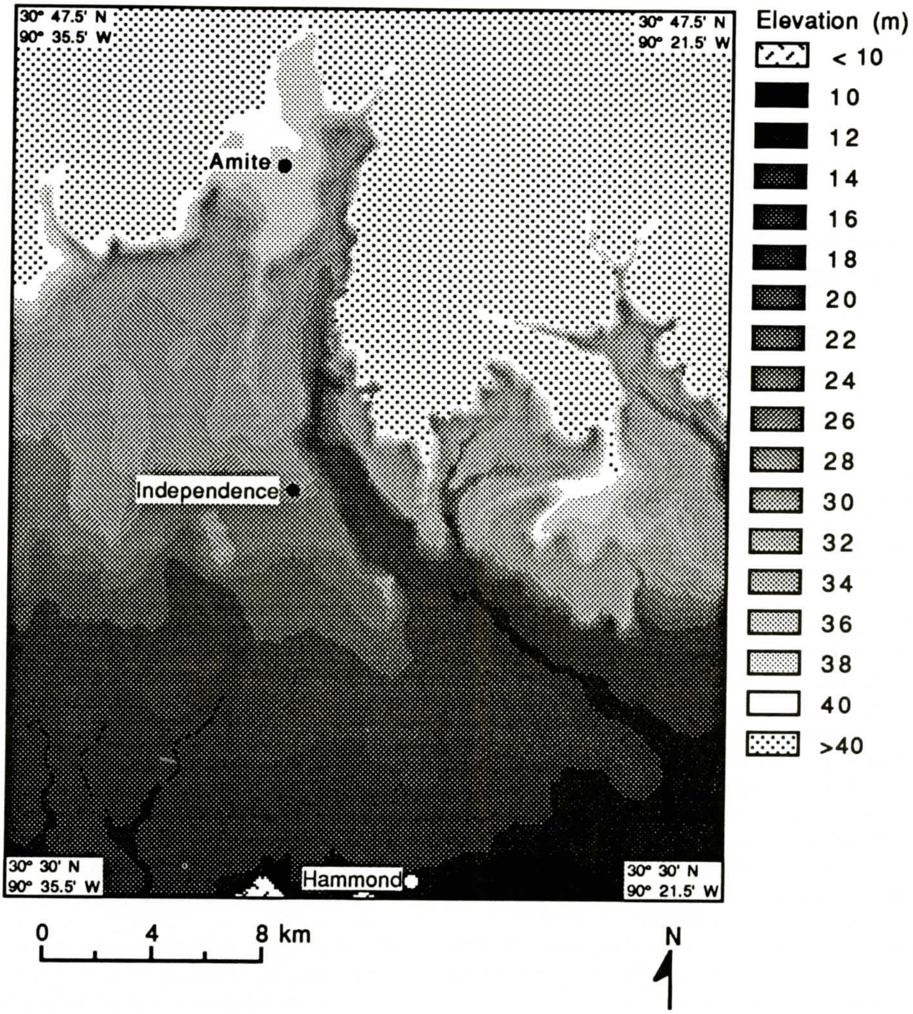

Figure 2. Map derived from a 1:250,000 Digital Elevation Model showing landscape elevations between 10 and 40 m (dominantly of late Pleistocene age) in west-central Tangipahoa Parish, Louisiana. The map shows several isolated areas of high relative elevation, indicating that erosion is an important aspect of regional landscape evolution.

some high areas that are not apparent on the 1:24,000 maps. Considering these limitations, a map of the study area was produced from this source (Fig. 2). From a topographic perspective, there are several connected and isolated areas of high relative elevation shown on these maps. A large component of the variability in elevation is erosional, whereby topographic remnants become isolated from the larger contiguous areas of the same geologic unit. More work is recommended using larger scales, such as 1:24,000, which would clearly show the sand hills and features of a similar size.

Hillslope studies elsewhere have shown spatial variability in sediment sources, transport, and processes of erosion (Young and Saunders, 1986), but little work of this nature has been conducted in the Gulf Coastal Plain. An intensive field research effort in the study area is desperately needed. Specific attention to spatial sampling design is important, and the use of geochronologic methods, where feasible, should be used to examine processes over long time scales. Such efforts, perhaps assisted by digital models, will provide further answers concerning landscape development in the region.

ACKNOWLEDGEMENTS

Thanks to David Fann for producing the Digital Elevation Model diagram and Mark McLean for modification of the geologic map showing inliers.

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